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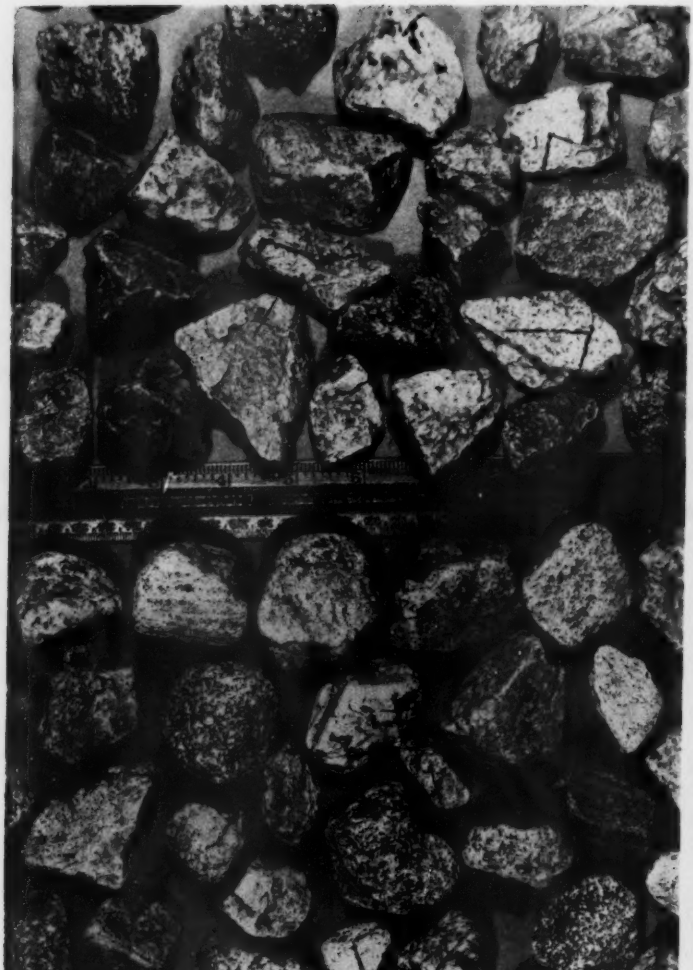

VARIATION IN ROUNDNESS OF SAN GABRIEL PEBBLES

Shown by contrasting sample O-2 (above ruler) with sample O-14 (below ruler). These samples are 5.5 miles apart in the field.

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FLOOD GRAVEL OF SAN GABRIEL CANYON, CALIFORNIA

BY W. C. KRUMBEIN

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ABSTRACT

Fifteen channel samples and six oriented samples were collected from the flood gravel left during the great flood of March 1938, in San Gabriel Canyon, California. The samples extended along 7 miles of the canyon, from above Dam No. 1 to Sycamore Flat, near the headwaters. The gravel was analyzed for size, shape, roundness, lithology, and fabric. Size, degree of sorting, shape, and lithology showed mainly nonsystematic variations along the valley; roundness and fabric showed systematic changes as a function of distance along the valley. Pebble imbrication changed from a downstream direction near the head of the canyon to typical upstream imbrication farther downstream.

The problem of evaluating the analytic data was complicated by the presence of several log jams along the valley, which imposed some limitations on the sampling process. The nonsystematic fluctuations of some of the sedimentary characteristics are attributed to similarity of depositional conditions initiated by the log jams.

An important part of the study is the development of a theory of pebble rounding, on the assumption that the rate of rounding is proportional to the difference between the initial roundness and the final roundness value which is approached by the pebbles in the given environment. The analysis yields a function of the type $P = P_o\ (1 - e^{-kx})$, where P is the roundness at any point along the stream, P_o is the final roundness value approached by the pebbles, k is a constant, and x is the distance along the stream. The observed data agree fairly well with the theory.

INTRODUCTION

One of the largest floods recorded in the Los Angeles region occurred on March 2, 1938. Continuous rainfall of average intensity for several days prepared the way for the major flood by thoroughly saturating the mantle rock. On the evening of March 1 the rainfall intensity abruptly increased, so that most of the rain ran directly into the streams and produced the major flood of March 2. The extent of the flood, the damage caused, and various engineering consequences are discussed by Burke (1939).

The present paper is concerned with the deposits left by the flood waters in San Gabriel Canyon, one of the larger valleys affected. San Gabriel River emerges from the San Gabriel Mountains near the town of Azusa, California. The valley is about 20 miles long and trends generally north-south through the mountains. About halfway along its course the North and West forks combine to form the major stream, and several miles farther south the East Fork adds its waters (Fig. 1).

Three dams have been built along the valley. Morris Dam, a concrete structure which impounds the water supply for Pasadena, is located a few miles above the mouth of the canyon. Dam No. 1, several miles farther upstream, is an earth structure which serves in part as a "boulder

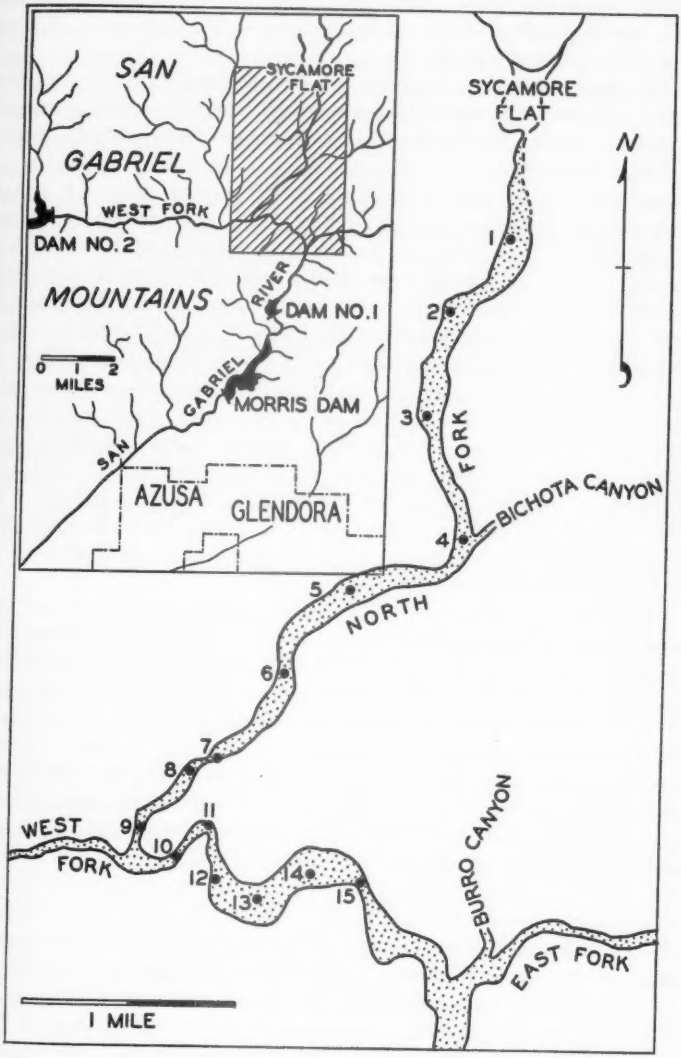

FIGURE 1.—*Map of portion of San Gabriel Canyon*
Inset shows regional setting of the Canyon.

strainer" for Morris Dam. Dam No. 2, also earthen, lies along the West Fork, about 6 miles above its junction with North Fork. All three dams held during the flood.

Flood waters quickly accumulated in San Gabriel Canyon. The crest moved downstream, sweeping out roads and bridges, as well as summer homes and camps. Within less than 24 hours the water reached spillway level in the reservoir.

Part of the valley only is concerned in the present study. It extends from approximately the junction of East Fork, upstream to and beyond the junction of West Fork, to the upper headwaters of North Fork at Sycamore Flat. In Figure 1 the dotted area represents approximately the width of the flood deposits. During the height of the flood, and for some time after, the lake behind Dam No. 1 extended upstream to a point about 1½ miles north of the junction with East Fork. When the flood waters diminished, the bottom of the portion of the valley studied was covered with a broad expanse of sandy gravel. Subsequently, and especially after the lake was drained, normal stream processes cut a shallow channel through the flood deposits, exposing the beds and affording an excellent section for 7 miles along the main canyon. The writer visited the canyon in July 1939, by which time the roads were largely reconstructed, but the major features and continuity of the flood deposits were essentially intact.

The writer is indebted to numerous geologists who have been interested in the problems raised by the study. Messrs. Carl Brown, Ian Campbell, James Gilluly, and G. H. Otto have visted the canyon and examined the field relations; the two latter have read the manuscript. Dr. L. G. Straub of the Institute of Technology of the University of Minnesota made valuable suggestions regarding the behavior of sedimentary particles during transport. Dr. F. J. Pettijohn has offered suggestions in the treatment of the data, and Dr. Carl Eckart of the Department of Physics of the University of Chicago proved helpful in the mathematical treatment of a theory of pebble rounding.

PURPOSES OF THE STUDY

The excellent exposures along San Gabriel Canyon and the known origin of the deposits afforded a unique opportunity of studying the characteristics of flood sediments. In contrast to normal streams, the large volumes, high velocities, and enormous loads of streams in flood may be expected to accelerate such processes as abrasion and rounding; in addition, the essentially continuous beds left behind presumably simplify the problem of collecting samples which are correlative along the entire valley.

A study of flood deposits may be expected not only to supply quantitative data on the fundamental characteristics of the deposits (size, shape, lithology, and fabric) but it may also afford critical data for comparing flood and normal stream deposits; similarities and differences disclosed may prove of value in interpreting ancient fluvial sediments.

The present study is confined to a 7-mile stretch of flood deposits along the main valley from a point slightly downstream from Sycamore Flat to the upper limit of the slackwater deposits behind Dam No. 1. The paper describes how size, shape, roundness, lithology, and fabric of the gravel vary as functions of distance along the valley and insofar as possible it relates these changes to the physical processes which occurred during the flood.

It is planned to extend the investigation to additional details about the deposits and to compare them with similar data from other valleys in the same region. In addition, the larger problem of the degree to which quantitative data reflect field relations will be examined, in terms of the sampling problems involved. The sampling of sediments is at best a difficult procedure, and an important purpose of the complete study is to evaluate sampling procedures in the light of the data obtained.

COLLECTION OF SAMPLES

FIELD WORK

In preparation for the collection of the samples and the subsequent evaluation of the data, an attempt was made to reconstruct conditions during the flood, drawing upon the present field relations of the deposits, reports of direct observations made during the flood, and evaluations made by observers after the worst stages had passed. In this connection the writer is particularly indebted to Mr. G. H. Otto of the Soil Conservation Service, California Institute of Technology, Pasadena. Mr. Otto gave the writer his continued co-operation and aid both during the field sampling and in the subsequent evaluation of the data. Otto recently developed the concept of the sedimentation unit for field sampling (Otto, 1938), and during the collection of the samples this concept was subjected to fairly rigorous tests. The sedimentation unit at any sampling point is essentially that portion of the material which was deposited under statistically constant physical conditions. The unit may consist of a single bed or a group of beds. Its recognition in the field depends upon an evaluation of the conditions of deposition over the area to be sampled, and once the unit is chosen it is followed rigorously.

After the quantitative data on size, sorting, shape, and roundness were available, Messrs. C. Brown. I. Campbell, J. Gilluly, and G. H.

Otto revisited the field and placed their findings at the disposal of the writer. The general field relations discussed here are a summary of the salient features obtained by the writer and Mr. Otto, plus the findings of the later field party.

A brief summary of conditions at the initiation of the flood has already been given. The sharp increase of rainfall intensity on an already saturated mantle resulted in a high proportion of surface runoff; moreover, the lubricated mantle moved bodily down the steeper slopes, as well as in the steeper rivulets, and numerous landslide scars in the upper reaches of the valley attest the importance of such debris as a source of the sediments. It is estimated that from half to three-quarters of the load carried by the flood came from such sources at or above Sycamore Flat. Supporting data were found in the absence of marked scour and fill at the junctions of tributaries below Sycamore Flat, which suggests that these tributaries did not bring in a major portion of the load. Earlier deposits along the floor of the valley were in many places completely reworked, and locally high terraces of earlier gravels contributed material, as shown by fresh exposures.

It seems likely that both erosion (transport) and aggradation went on simultaneously during the rising and ebbing stages of the flood. The interplay between the two varied from time to time and from place to place, depending upon such factors as gradient and channel width. Numerous log jams at intervals along the valley complicated matters. The present field relations show immediately upstream from the jams a coarse gravel which grades into well-bedded medium gravel farther upstream. Still farther upstream, at the base of the next higher log jam, is a region of scour with a lag sediment of coarse boulders. The flood debris is thus to some extent a series of partly overlapping deposits related to the log jams. In the main the present deposits probably represent flood ebb accumulations, inasmuch as pre-crest jams would be torn out by the higher waters.

Detailed examination of the log jam accumulations indicated the general presence (in the well-bedded portions) of three phases of sedimentation. The lower several feet were coarse and seemed to be less homogeneous than the intermediate layers, which are several feet thick. At the very top of the gravel scattered cobbles and boulders were noticeably coarser than the average sizes in the well-bedded intermediate section. It was concluded that three sedimentation units were present and that at least the intermediate gravels represent deposition from comparatively slack water above log jams. Some of the coarse lower material may consist of boulder bars formed behind fixed obstructions in the channel floor. In many cases, however, this lower material prob-

ably represents debris which dropped immediately upon initiation of a log jam. A considerable portion of finer material near the stream bed was entrapped with it, resulting in a poorly sorted deposit. This lower sedimentation unit may extend indefinitely upstream because of the slackwater effect produced by the temporary damming. Meanwhile, heavy deposition continued in the ponded areas near the log jam, the debris being supplied by the partially "cleared" waters which continued to pour in from upstream. In this manner the intermediate well-bedded sedimentation unit accumulated. The upbuilding of the beds forced the stream to cut laterally around the jam unless the jam itself was partially undermined by the clear water pouring over its edge. In either case a channel lower than the surface of the deposit was soon developed. As the water collected into this new channel some of the topmost layers were swept downstream, leaving behind a lag deposit of cobbles and boulders, which is the topmost sedimentation unit.

On the basis of these observations, samples were taken from the intermediate, relatively well-bedded gravel only. In terms of sedimentation units, it is believed that this material is correlative from one log jam to the next, not in time directly but strictly in terms of flood stage (*i.e.*, ebb) and conditions of deposition. It was hoped that, as a first approximation, this sampling choice would afford comparable data which could be used to evaluate the changes in sediment characteristics along the valley. Some of the complexities which arose will be discussed in a later section.

After the sampling unit was chosen, subsequent samples were collected insofar as feasible from comparable sites. Two types of samples were collected. The first were channel samples taken through the entire intermediate sedimentation unit. These were collected by means of a special metal scoop designed by Mr. Otto. The channel samples averaged 15 pounds, believed to be large enough to avoid significant errors in the coarse sizes. These samples, 15 in all, were collected at approximately half-mile intervals along the valley. The second type consisted of samples of 100 oriented pebbles each. These pebbles were collected with an orienting frame in a manner described elsewhere by the writer (Krumbein, 1939a). The process consists of marking a right angle on the pebble while it is still in the outcrop. This right angle is distorted into curved lines when the pebble is viewed from any position except directly in front. As a result, the field orientation of the pebble may be reconstructed in the laboratory with a goniometer.

The oriented pebbles ranged from about 15 to 70 mm. in long diameter. These samples were collected at intervals of approximately a mile along the valley. Six samples were taken and were used for a study of their

orientation (fabric) and for shape studies. The channel samples were used for studies of size, sorting, and lithology.

The sampling points are indicated in Figure 1. The choice of the particular stretch of valley indicated in that figure was governed by

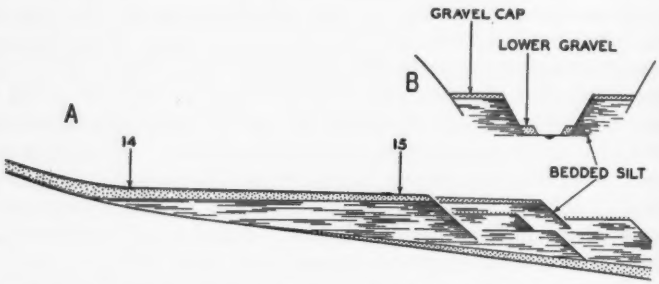

FIGURE 2.—*Relations between gravel and silt along lower part of valley*
(A) Longitudinal section in vicinity of samples 14 and 15. (B) Cross section of valley showing relation of gravel to silt.

several factors, of which accessibility was one. More important was the fact that above sample 1 the deposits were discontinuous, extremely heterogeneous, with ill-defined bedding and very poor sorting. The first sample was accordingly collected at about the point where bedding became prominent, although the gravel seemed to be poorly sorted. At the lower end of the sample series arose a question of the exact identity of the sedimentation unit. The unit was traceable to sample 14 (Fig. 1) but at about that point it was found that the gravel was underlain by slackwater sand and silt. Downstream from sample 14 the silts increased in thickness, and a single layer of gravel overlay them, with a very thin veneer of large cobbles. The silt and gravel were separated by a sharp transition, and it appeared unquestionable that the gravel belonged to the intermediate sedimentation unit.

Figure 2A is a sketch of the relations along the stream in the vicinity of samples 14 and 15. The gravel cap on the main silt terrace thins downstream and becomes more sandy. In addition, terraces at several levels develop downstream. These terraces carry a cap of gravel, and the several levels appear to be related to stages of the lake level during the dying stages of the flood. In view of these features it was decided to take only a single sample downstream from 14, and that from the highest terrace, to avoid extrapolating the sedimentation unit too far. During the year of normal stream action which followed the flood, a channel was cut through the several deposits, and some gravel was

moved downstream. As a consequence low terraces of gravel also flank the eroded silt beds. These relations are shown in Figure 2B. The cross section was taken just above the junction with East Fork, where the silt terraces are well developed. The section shows that the gravel in the channel abuts against the eroded silts and is thus later.

DESCRIPTION OF SAMPLES

Sample 1. Channel sample, 0.6 mile below Sycamore Flat, near center of valley flat. The gravel is poorly bedded and "dirty." Sedimentation unit 3 feet thick.

Sample 2. Channel sample, 0.5 mile downstream from 1, in west center of wide valley flat. Gravel poorly bedded and "dirty." Very large boulders (10 feet diameter) scattered on flat.

Sample 0-2. Oriented sample of 100 pebbles, collected at site of sample 2. Valley trends N. 53° E.

Sample 3. Channel sample 0.5 mile downstream from 2, in western part of wide valley flat. Bedding noticeably better than farther upstream. Thickness of unit 2.5 feet.

Sample 4. Channel sample 0.6 mile downstream from 3, near center of wide flood-plain. Well-bedded, with some openwork layers. About ¼ mile upstream is a large cut-bank of earlier gravel. Large boulders still numerous; a 12-foot boulder about 20 feet west of sample.

Sample 0-4. Oriented sample at site of sample 4. Valley trends N. 12° W.

Sample 5. Channel sample 0.7 mile downstream from 4, near center of wide flat. Large cut-bank of earlier gravel ¼ mile upstream from sample. Bedding fair.

Sample 6. Channel sample 0.6 mile downstream from 5, near center of wide flat. Bedding not well developed. East of the site is a large cut-bank of earlier gravel. Scattered boulders lie on surface of flat, largest about 8 feet in diameter. Numerous 1-foot boulders in vicinity.

Sample 0-6. Oriented sample at site of sample 6. Valley trends N. 1° E.

Sample 7. Channel sample 0.5 mile downstream from 6, along eastern wall. This site is at the upper end of a narrows. Gravel well-bedded and imbricated.

Sample 8. Channel sample about 0.2 mile downstream from 7, near center of flat just below narrows. The gravel is well-bedded and imbricated. Sedimentation unit 2 feet thick.

Sample 0-8. Oriented sample at site of sample 8. Valley trends N. 69° E.

Sample 9. Channel sample 0.4 mile downstream from 8, along eastern wall of valley. The gravel is well-bedded and imbricated. Sedimentation unit is 2 feet thick.

Sample 10. Channel sample 0.5 mile downstream from 9, near center of wide flat. The sampling site is 0.4 mile downstream from the junction with West Fork. The larger boulders scattered on the flat here are more numerous than at 9, suggesting coarser material from West Fork.

Sample 11. Channel sample 0.3 mile downstream from 10, from a bar-like feature in the acute angle of a sharp bend in the valley. The gravel is well-bedded and imbricated.

Sample 0-11. Oriented sample from site of sample 11. The valley upstream trends N. 31° E.; downstream the trend is N. 7° W. The bar trends N. 64° W.

Sample 12. Channel sample 0.3 mile downstream from 11, near western wall of valley. Gravel is well-bedded. Large boulders are scarce, but cobbles are abundant on the flat.

Sample 13. Channel sample 0.4 mile downstream from 12, near center of wide flat. The sampling site is opposite the mouth of a tributary at Camp Rincon. The gravel is well-bedded. Cobbles are common on the flat.

Sample 14. Channel sample 0.3 mile downstream from 13, near center of wide flat. Sedimentation unit about 2.5 feet thick. Gravel is well-bedded and imbricated. Bedded silt is exposed below the gravel.

Sample 0-14. Oriented sample from site of sample 14. Valley trends N. 76° E.

Sample 15. Channel sample 0.4 mile downstream from 14, in gravel cap above bedded silt. The gravel is very sandy; the sedimentation unit is about 2 feet thick.

LABORATORY ANALYSIS OF THE SAMPLES

GENERAL STATEMENT

The channel samples were subjected to mechanical analysis by sieving and to lithological counts. The oriented samples were studied first for the fabric of the deposit, and the same samples were then used for shape and roundness studies. These four attributes, as well as surface texture, constitute the essential textural and compositional features of sediments. Surface texture has not yet been placed on a quantitative basis.

MECHANICAL ANALYSIS

All the samples had material ranging from about 64 mm. diameter to $\frac{1}{16}$ mm. diameter, plus a small amount in the subsieve sizes. This range includes 11 Wentworth grades. For statistical analysis it is desirable that the sieving process should yield at least six grades, with not too large a percentage of material in any single grade. The data so obtained are suitable both for cumulative curves and for the computation of moments. On the basis of this criterion, sieves at Wentworth intervals were used for all analyses.

The entire sample (average weight 6.5 kilograms) was used for the coarser sizes down to 4 mm. Beyond this the sample was quartered into a smaller sample suitable for the sand ranges. The data were combined and expressed as weight percentages. Table 1 lists the mechanical composition of the 15 channel samples.

Inspection of the table indicates that no particular grade dominates the distributions. In general the material is poorly sorted in the aggregate, because of the secondary modes. These at least suggest that more than one frequency distribution is involved in the deposit. The bimodal tendency becomes more pronounced downstream, especially in sample 15. The intimate mingling of sand and gravel in the same sedimentation unit renders it unlikely that the material represents two episodes of deposition, such as an openwork gravel with the later infiltration of sand. This conclusion is supported by the fact that most of the

exposures contain pebbles completely surrounded by sand; well-developed openwork seams are rare. If it is assumed that the gravel constituted a traction load and the sand a suspension load, it seems probable that deposition from both loads was essentially simultaneous.

TABLE 1.—*Size composition of San Gabriel flood deposits*
(Expressed as weight percentages)

Sample	Grade Limits in mm.											
	128–64	64–32	32–16	16–8	8–4	4–2	2–1	1–½	½–¼	¼–⅛	⅛–1/16	Pan –1/16
1	.....	11.8	18.9	16.1	13.4	8.5	8.3	8.8	7.8	4.0	1.5	0.9
2	11.8	11.1	23.8	14.9	10.6	7.1	5.7	4.9	3.7	2.9	1.8	1.7
3	.....	9.6	12.1	21.1	16.2	11.3	10.0	9.7	6.5	2.6	0.6	0.3
4	.....	4.5	24.1	18.2	14.4	9.7	7.3	7.5	7.5	4.7	1.4	0.7
5	.....	18.3	21.4	12.7	9.6	8.6	8.2	8.1	7.1	4.0	1.4	0.6
6	.....	35.4	16.9	16.1	9.6	5.8	3.8	3.7	4.5	2.7	1.0	0.5
7	.....	15.0	12.7	14.8	12.4	11.7	11.1	10.8	7.4	3.0	0.8	0.3
8	.....	27.1	23.8	16.1	7.1	3.4	3.4	6.3	7.1	3.9	1.3	0.5
9	.....	8.6	24.4	23.7	13.8	6.6	4.8	4.9	6.7	4.8	1.3	0.4
10	8.3	15.8	10.6	11.6	8.4	8.4	13.6	12.0	7.2	3.0	0.8	0.3
11	.....	29.0	24.3	15.0	7.3	4.4	3.2	4.7	8.4	2.9	0.6	0.2
12	.....	13.2	15.8	17.0	14.8	11.8	12.3	8.7	4.3	1.5	0.4	0.2
13	.....	15.1	19.1	12.3	9.7	9.8	11.7	11.7	6.2	3.0	1.0	0.4
14	.....	18.9	16.8	17.1	10.0	8.2	9.6	9.4	6.4	2.6	0.7	0.3
15	.....	11.1	12.3	9.6	8.5	8.7	11.5	14.9	15.7	5.5	1.6	0.6

SHAPE ANALYSIS

The sets of pebbles collected for orientation analysis were used for studies of shape. The pebbles range from about 15 to 70 mm. in long diameter. Fifty pebbles were used from each sample; the larger pebbles, most of them in the range 32 to 64 mm., were chosen as a standard. Hence the interpretations of shape in this paper are confined to pebbles in that range.

Shape of pebbles, as the term is used here, refers to their sphericity as defined by Wadell (1932). The sphericity is the ratio of the nominal radius of the pebble to the radius of the circumscribed sphere. The nominal radius is the radius of a sphere having the same volume as the pebble. On this basis, sphericity, Ψ, is defined as

$$\Psi = r_n/r_s \quad \ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots (1)$$

where r_n is the nominal radius and r_s is the radius of the circumscribed sphere.

Details of sphericity measurement are given by Krumbein and Pettijohn (1938, Chapter 11). Essentially the procedure involves determining

the longest diameter and the volume of the pebble. The latter is found by displacement in water. With these data a nomogram may be used to compute the sphericity. Sphericity varies from 0 to 1, the latter associated with true spheres.

The sphericity data were assembled into 0.05 Ψ classes and expressed as number percentages (Table 2). The data spread over a varying

TABLE 2.—*Sphericity distributions of San Gabriel gravel*
(Expressed in number percentages)

Sample	0.50–.54	0.55–.59	0.60–.64	0.65–.69	0.70–.74	0.75–.79	0.80–.84	0.85–.89	0.90–.94
0–2	2	10	30	36	14	8	.....	.....	.....
0–4	2	12	12	36	18	8	8	2	2
0–6	.....	12	18	18	32	16	4	.....	.....
0–8	4	24	14	16	26	12	4	.....	.....
0–11	6	8	20	24	18	18	2	4	.....
0–14	.....	.....	14	38	18	24	6	.....	.....

number of classes, and both unimodal and bimodal distributions occur. Sample 0-14 is in rather striking contrast to the other samples, because despite its bimodal nature its range of sphericity is confined between 0.60 and 0.85.

ROUNDNESS ANALYSIS

The same 50 pebbles from each set used to measure sphericity were studied for roundness. Hence for roundness as well as shape the interpretations are confined approximately to the 32-64 mm. range. A qualitative examination was made of the roundness of small particles in the channel samples, and those data will be introduced later.

Roundness may be expressed in several manners. Wadell's (1932) method is followed in its essentials. Wadell defined roundness as

$$P = \frac{\Sigma\left(\frac{r_i}{R}\right)}{N} \qquad (2)$$

where P is the numerical measure of the roundness, r_i is the radius of curvature of any particular "corner" or edge on the pebble, R is the radius of the inscribed circle within the pebble, and N is the number of corners or edges measured. The sign Σ refers to a summation process. The measurements are made by projecting the image of the pebble to a standard magnification of 7 cm. long dimension and using the image to measure the radii. (For details, *see* Krumbein and Pettijohn, 1938, Chapter 11.)

The roundness value of a pebble may be visualized geometrically very readily by re-expressing equation (2) as

$$P = \frac{\Sigma\left(\frac{r_i}{N}\right)}{R} \quad \ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots \quad (3)$$

This expression yields the same numerical values as equation (2), but it indicates that roundness may be considered as the ratio of the arithmetic mean of the corner radii to the radius of the inscribed circle. It is thus a statistical measure of the degree to which the "average" corner or edge of the pebble approaches the curvature of the inscribed circle. Roundness varies from 0 to 1; it is obvious from equation (3) that, if the pebble edges are sharp, the average radius of the corners will be much less than R, so that the roundness will approach zero. With rounding, the ratio approaches 1. As will be shown, this geometrical picture of roundness affords a simple basis for developing an analytical theory of pebble rounding.

TABLE 3.—*Roundness distributions of San Gabriel gravel*
(Expressed in number percentages)

Sample	0.10–.14	.15–.19	.20–.24	.25–.29	.30–.34	.35–.39	.40–.44	.45–.49	.50–.54	.55–.59	.60–.64	.65–.70
0-2	2	14	22	30	18	8	6	.....	.....	.....	.....	.....
0-4	.....	.....	4	8	22	30	10	12	6	2	.....	.....
0-6	.....	.....	2	8	16	26	20	14	10	4	.....	.....
0-8	.....	.....	.....	6	12	28	24	18	6	2	4	.....
0-11	.....	.....	.....	2	10	22	28	18	6	8	4	2
0-14	.....	.....	.....	4	10	18	26	24	8	6	2	2

The roundness data of the samples are given in Table 3, arranged in 0.05 P classes. In contrast to the sphericity distributions of Table 2, the roundness distributions show a progressive change in their modes from approximately 0.275 to 0.425.

ORIENTATION ANALYSIS (FABRIC)

The preferred orientation of a pebble in a deposit may be expressed in various ways. Most commonly the dip direction and angle of dip of the longest axis are used. Elsewhere the writer (Krumbein, 1939a; 1939b) discusses in detail the use of the long axis as one index and the pole of the maximum projection plane of the pebble as the second index. The long axis of the pebble is measured with a sliding block (Krumbein and Pettijohn, 1938, p. 145), and the pole of the plane is located by means of a photo-electric cell.

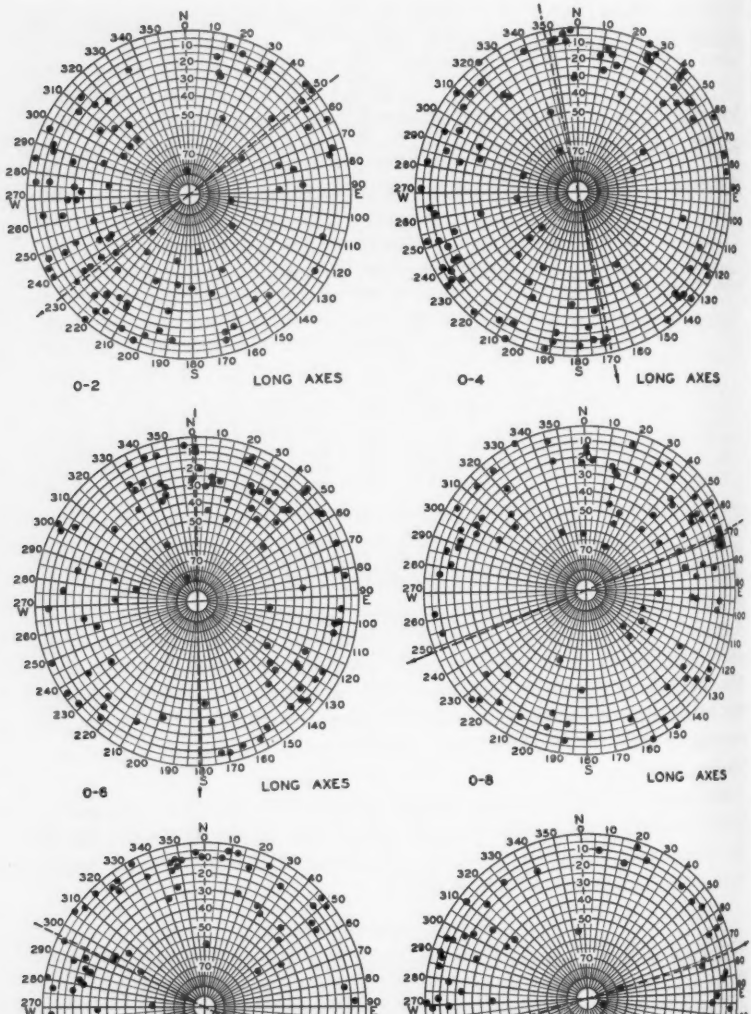

FIGURE 3.—*Polar co-ordinate charts of pebble orientation*
The sample numbers are shown in the lower left of each diagram.

Although the writer emphasized the need of two reference lines for the complete fabric of a pebble deposit, in special cases one or the other is adequate for a first approximation. Stream pebbles, for instance, usually lie with their maximum planes gently inclined, and with their long axes dipping upstream. Inasmuch as the long axis lies in or parallel to the maximum plane, the long axis alone may in ordinary circumstances be used to study imbrication. By observing the pebbles as they are measured, the need for more detailed analysis may be evaluated. In the present study only the long axes were measured. The collected pebbles, each with its orientation mark, are placed in a sliding block, and the points of emergence of the long axis are marked with lacquer. The pebbles are then mounted in a two-circle goniometer, and the dip direction and angle of dip of the long axis read by using the horizontal and vertical circles. The dip direction is expressed as an azimuth measured clockwise from the north point.

In assembling the orientation data, it is more convenient to plot the dip direction and dip on polar co-ordinate paper than to list them in tables. Figure 3 shows the observational data for the six samples. A given point has the co-ordinates dip direction (measured along the circumference) and angle of dip (measured inward along a radius). Thus the point in 0-2 (Fig. 3) with azimuth 335° and dip 16° means the long axis is dipping downward 16° along a line striking N. 25° W.

The arrows drawn on the diagrams indicate the downstream trend of the valley at the sampling point. Inasmuch as the flood waters covered the valley bottom it was assumed that the direction of stream flow was parallel to the valley trend. It is apparent even from the polar co-ordinate charts that there is a preferred orientation with respect to the arrow in most cases. An exception in the drawing of the arrow is sample 0-11 (Fig. 3), in which the sample was collected from a sharp bend in the valley. The arrow in that diagram is drawn parallel to the bar-like deposit from which the pebbles were collected.

LITHOLOGICAL ANALYSIS

No study was made of the areal geology near the canyon, but an approximate pebble count was made of the samples to see whether the lithology remained fairly constant or whether any pronounced changes occurred downstream. For this purpose approximately 100 pebbles of diameter 16 to 64 mm. were counted in each sample. Most of the pebbles were granitic, but for classification purposes four broad groups were chosen, partly on the basis of structure and composition, inasmuch as there were complete transitions from one group to another. The final subdivision included (1) granitoid rocks (granite, pegmatite, and gneissic

granite); (2) dioritic rocks (rocks of intermediate composition); (3) fine-grained dark schistose rocks (amphibole schists and related rocks); and (4) miscellaneous rock types (felsite porphyry, greenstone, and some metamorphic rocks). The line between (1) and (2) was not always easily chosen. The data of the pebble counts are given in Table 4.

TABLE 4.—*Lithological composition of San Gabriel gravel*
(Expressed as number percentages)

Sample	Granitoid pebbles	Dioritic pebbles	Schistose pebbles	Miscellaneous pebbles
1	80.0	5.5	14.5	.....
2	88.9	6.5	9.6	.....
3	79.8	6.2	11.0	.....
4	83.9	5.8	10.3	.....
5	83.5	9.7	4.9	1.9
6	86.3	7.9	4.5	1.3
7	83.9	8.8	7.3	.....
8	87.5	8.3	4.2	.....
9	83.2	8.4	8.4	.....
10	67.3	9.6	13.5	9.6
11	68.5	16.6	8.6	6.3
12	75.0	11.7	8.3	5.0
13	68.3	10.6	15.3	5.8
14	75.9	10.3	9.2	4.6
15	73.4	8.2	12.3	6.1

STATISTICAL TREATMENT OF THE DATA

The present study is concerned with the functional relationships between sediment characteristics and position along the valley. Such a study can be more effective if the data are quantitative and in statistical form. The statistical values summarize the data and afford a basis for plotting the values of the several characteristics against distance. The resulting curves represent the changes in the attributes as functions of distance from source. A study of these changes may suggest physical processes operative during the flood, and they provide quantitative tests of various theories which may be set up to account for the observed changes.

Of the many kinds of statistical measurements, moment measures have certain advantages. Moment analysis is applicable whether the data are arranged in arithmetical or logarithmic classes; further, the geometrical significance of the measures is always the same, whether applied to size, shape, roundness, lithology, or fabric. The analytical data of this study are summarized in terms of the first two moments of the distributions. The first moment is the arithmetic mean of the frequency distribution in terms of the independent variable used. If the

TABLE 5.—*Statistical parameters of size data*

Sample	Phi Mean M_ϕ	Phi Standard Deviation σ_ϕ	Geometric Mean Diameter in mm. GM_ξ
1	−2.21	2.52	4.67
2	−3.08	2.68	8.45
3	−2.19	2.23	4.56
4	−2.16	2.40	4.46
5	−2.50	2.59	5.66
6	−3.40	2.40	10.50
7	−2.17	2.41	4.50
8	−3.07	2.60	8.40
9	−2.56	2.37	5.90
10	−2.33	2.66	5.02
11	−3.23	2.48	9.37
12	−2.57	2.17	5.95
13	−2.29	2.48	4.90
14	−2.57	2.44	5.92
15	−1.47	2.64	2.76

TABLE 6.—*Statistical parameters of shape and roundness data*

Sample	Shape		Roundness	
	Mean sphericity	Standard deviation	Mean roundness	Standard deviation
0–2	0.66	0.056	0.28	0.071
0–4	0.69	0.082	0.38	0.077
0–6	0.69	0.072	0.40	0.082
0–8	0.67	0.080	0.42	0.079
0–11	0.69	0.083	0.44	0.087
0–14	0.71	0.058	0.44	0.084

TABLE 7.—*Statistical parameters of orientation data*

Sample	Azimuth		Angle of Dip	
	Mean azimuth	Standard deviation	Mean dip	Standard deviation
0–2	64.0°	44.0°	27.3°	17.4°
0–4	19.2	46.8	19.3	16.4
0–6	165.4	47.6	23.3	15.2
0–8	65.2	48.5	24.1	17.2
0–11	97.8	49.7	20.2	14.7
0–14	83.0	48.2	21.1	16.4

data are arranged on an arithmetic scale (as in shape, roundness, fabric), the first moment is the arithmetic mean; if the data are logarithmic (as in size), the first moment is the logarithmic mean. In either case the geometric significance of the first moment is the same. It is the center of gravity of the distribution, and hence lies in the central portion of the distribution.

The second moment, or more properly its square root, is σ, the standard deviation. This also may be computed either for arithmetic or logarithmic data. In either case it is a measure of the spread of the data about its center of gravity. In a symmetrical bell-shaped curve the range (mean $\pm\sigma$) includes about 68 per cent of the distribution. In any distribution it is a measure of the width of the main central part of the distribution in terms of the scale units used. Details of computing the moments, and additional information on their geometrical significance, are given by Krumbein and Pettijohn (1938, Chapters 8 and 9).

The size data of Table 1 were summarized as logarithmic moments and are shown in Table 5. The phi mean (M_ϕ) is the center of gravity of the logarithmic distribution, and the phi standard deviation (σ_ϕ) is the corresponding measure of spread. The diameter equivalent of the phi mean, the geometric mean diameter in mm. (GM_ξ), is also given for those who prefer that measure (Krumbein, 1936).

Table 6 includes the first two moments of the shape and roundness distributions. These moments are arithmetic inasmuch as the data are fairly symmetrical in their original equal-interval classes. Table 7 includes the first two moments of the orientation data. These latter data deserve some discussion. The azimuthal distributions consider only the azimuth of the pebble axes in terms of combined classes over a range of 180°. (*See* Krumbein, 1939a, for a discussion of the theory.) The mean azimuth can be related to the trend of the valley (direction of flood flow), and the azimuthal standard deviation is a rough measure of the degree of preferred azimuthal orientation. The dip distributions of the orientation data consider the angle of dip of the axes but not the dip direction. In addition to the numerical orientation data, standard petrofabric diagrams will be used in the interpretation of fabric.

The lithological data will be used as given in Table 4, and no further statistical work was performed on them.

PRESENTATION OF THE DATA

SIZE DATA

Figure 4 shows the average size and the degree of sorting of the flood deposits as a function of distance along the valley. The phi mean, M_ϕ, and its corresponding geometric mean diameter in mm., GM_ξ, are

shown in the upper figure. The phi standard deviation, σ_ϕ, is shown in the lower graph. This parameter may be interpreted physically as a measure of the degree of sorting of the sediment. The direct geometrical meaning of σ_ϕ is that its value indicates half the number of Wentworth

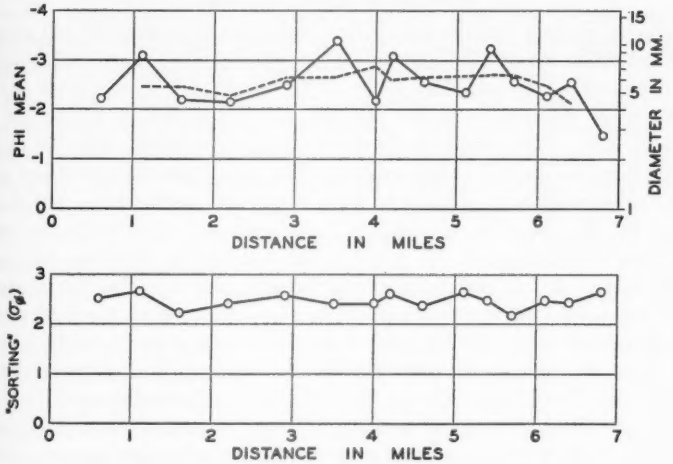

FIGURE 4.—*Mean size and degree of sorting as functions of distance*
Downstream from Sycamore Flat. Upper graph, phi mean and geometric mean diameter; lower figure, phi standard deviation of samples.

grades in the main central portion of the distribution. The value $\sigma_\phi = 2.5$, for example, indicates that the central part of the size distribution is spread over five Wentworth grades. Such a sediment is obviously poorly sorted in comparison, say, with the average beach gravel.

The behavior of the phi mean with distance is noticeably erratic. Except for sample 15 the values range widely, and apparently at random. This absence of a size trend is emphasized by the broken line in the graph, which represents a moving average of three observations at a time. (*See* Krumbein and Pettijohn, 1938, p. 198-199.) The moving average tends to smooth the larger variations, and its downward trend at the extreme right is occasioned largely by the significant drop in the flood deposits as a function of distance along the valley. The phi mean, last sample. It will be recalled that sample 15 was collected from the gravel cap on the silt terraces; it is the only sample which suggests a trend toward smaller sizes, which may mean that this sample does not belong to the same sedimentation unit as the others.

The degree of sorting, as measured by σ_ϕ, also shows an apparently random variation along the valley. In this connection, it is interesting to note that the value for the last sample on the right (15) is of the same magnitude as previous samples.

PROBLEM OF SAMPLING ERRORS

The apparently random fluctuation of the phi mean and the phi standard deviation raises the question of sampling errors. Any sedimentary study is subject to sampling errors of two kinds; the first relates to the position of the sample in the deposit, and the second is a question of sample size. The former is the field error of sampling, and the latter is the error of the individual sample. The use of the sedimentation unit for field sampling is believed to have materially reduced the positional error by the collection of samples from comparable parts of successive log jam accumulations. It was pointed out earlier that the log jams were interpreted as ebb flood accumulations, and that the intermediate sedimentation unit represents the immediate response of the load to the conditions set up by the log jam. Although the log jams probably did not all form simultaneously, nevertheless the deposits behind them should reflect, in a broad way at least, the ebb flood conditions along the valley.

Two or three exceptional samples in terms of position should be mentioned. Sample 7 was collected immediately above a very narrow portion of the valley (Fig. 1). In the same locality occur slackwater sands. Examination of Table 5 shows that although the geometric mean is noticeably smaller than the samples immediately above or below, it is not beyond the order of magnitude of such upstream samples as 1, 3, and 4, which are near the head of the series. Samples 14 and 15 were collected in the lower portion of the valley where log jams were not so conspicuous because of valley width. Nevertheless, sample 14 has a geometric mean comparable to its immediate predecessors, although that of sample 15 is appreciably smaller than any other. The field evidence of Figure 2 was accepted, however, as indicating the identity of the sedimentation unit of samples 14 and 15. Furthermore, the phi standard deviations of all samples are of almost the same order of magnitude, indicating no great fluctuations of sorting conditions anywhere along the valley.

The several observers who visited the field agreed to the essential continuity of the deposits sampled, although some doubt was expressed about the consistency of sample 7, and the writer has pointed out the possible exceptional nature of sample 15. On the whole it seems safe to conclude that most of the samples are consistent with respect to identity of flood conditions at the time of their formation.

The problem of individual sample size is perhaps the more difficult of the two sampling problems. The samples in the present study are channel samples extending from the bottom to the top of the intermediate sedimentation unit. They were collected with a metal scoop which prevented the loss of fine material. The samples average about 15 pounds each. The material in each sample ranges from 1/32 mm. to 64 mm. diameter in all samples, and to 128 mm. in two samples. The geometric mean particle size, however, ranges from 2.8 to 10.5 mm. diameter (Table 5).

The problem of sample size is complicated by the bimodal nature of the size frequency distributions (Table 1). In a normal distribution it is possible to estimate the probability of including larger pebbles in a given sample. A normal bell-shaped probability distribution has 16 per cent of its individuals larger than $(M_\phi + \sigma_\phi)$, 7 per cent larger than $(M_\phi + 1.5\sigma_\phi)$, and 2.5 per cent larger than $(M_\phi + 2\sigma_\phi)$ (Camp, 1931, Chapter 5). The present samples have 11 or 12 grades (Table 1) and an average value of 2.5 for σ_ϕ (Table 5). Owing to asymmetry and bimodality the larger pebbles in the present samples are between one and two standard deviations from the phi mean. This suggests that the probability of including additional large pebbles in a given sample may lie between (roughly) 20 and 10 per cent. In either case the center of gravity of the distribution would be shifted by a smaller percentage of its own value than these probabilities. Even if the net effect were an error of 10 per cent in the value of the phi mean, it would not seriously affect the nature of Figure 4. The writer accordingly believes that the samples are large enough to yield at least first approximations of the phi mean and phi standard deviation in terms of the particular sedimentation unit sampled.

It is evident that the question of sample size cannot be wholly settled with the data at hand, but fortunately that question and the larger question of the field error of sampling are amenable to quantitative attack. A graph could be prepared from experimental data to show the computed values of the phi moments as a function of sample size at any given sampling point. The computed values would change rapidly with small samples but would approach an asymptote with increase in sample size. From such data the error function could be evaluated. The field sampling error could be evaluated by choosing a set of samples on a grid laid over a log jam, and applying the probable error method (Krumbein, 1934), the theory of control (Otto, 1937), or the newer methods of analysis of variance. The writer hopes to include such studies in his further work on the San Gabriel deposits.

Similar problems of sample size are involved in shape, roundness, and orientation studies, as well as in pebble counts. Some unpublished data are available from work by Pettijohn and the writer, and it appears that

as long as statistical parameters are used in summarizing the data there is surprisingly little change in the computed values as the sample size increases from 50 to 100 pebbles. It was partly on this basis that relatively small samples were used for shape and roundness. In the latter, for example, from six to ten radii of curvature are measured per pebble, so

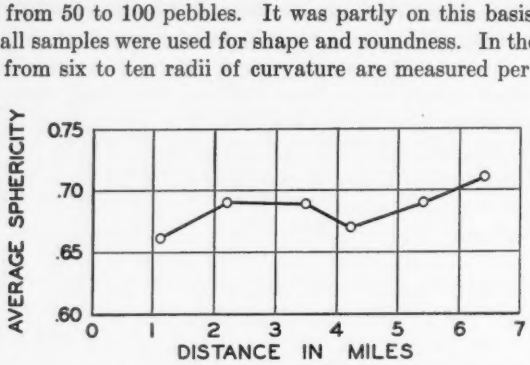

FIGURE 5.—*Mean sphericity as function of distance along the valley*

that each value is itself an average. Hence a further average of 50 pebbles yields a mean based on 300 to 500 individual observations.

The few data available on laboratory errors as distinct from sampling errors suggest that the laboratory error is much smaller. Earlier work on beach sands (Krumbein, 1934) showed that the field error in terms of the median diameter was 4.5 per cent, whereas the errors of splitting and mechanical analysis were less than 1 per cent. In the present interpretations it will accordingly be assumed that the laboratory errors are essentially negligible and that the sampling and field errors are small enough so that the data yield at least a first approximation to the characteristics of the deposits.

SHAPE, ROUNDNESS, FABRIC, AND LITHOLOGICAL DATA

Figure 5 shows the average sphericity as a function of distance along the valley. For purposes of comparison sphericity was determined on the same size pebbles as roundness. In order to have enough pebbles, the oriented samples alone were used. It would be desirable to have the sphericity data on all 15 samples, but they are not available. It is not apparent whether any definite trend exists, although there is a suggestion of a mean increase in sphericity; this increase at best is slight, however, the extreme values being 0.66 and 0.71.

The roundness data are shown in Figure 6. In contrast to previous figures, the mean roundness is well represented by a smooth curve. The progressive increase in roundness is striking when the samples are con-

trasted visually. Plate 1 shows the marked difference between pebbles in samples 0-2 (top) and 0-14 (bottom). The latter are approximately 5.5 miles downstream from the former, and the general appearance of the samples suggests that most of the pebbles have become more rounded, rather than that well-rounded pebbles from the upper parts of the valley

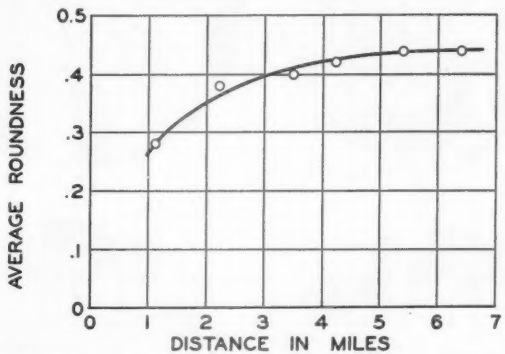

FIGURE 6.—*Mean roundness as function of distance along the valley*

were concentrated downstream by selective transportation. An examination of the same size pebbles in each of the 15 channel samples showed that the entire sequence is consistent with the conclusion that the pebbles progressively change in their roundness along the valley.

At first glance the roundness data and the size data contradict one another. If roundness increases as definitely as the data suggest, why does not the average size decrease in accordance with the wear implied by the increased roundness? The problem is puzzling, but an attempt will be made in the next section to reconcile the data.

Two methods of presenting the orientation data are used. Figure 7 shows six petrofabric diagrams of the long axes of the pebbles. In each diagram the vertical arrow indicates downstream direction. The diagrams were made from 100 observations each (Fig. 3) by the graphic methods described by Knopf and Ingerson (1938). The petrofabric diagrams show a striking sequence. In Figure 7, sample 0-2 (collected at site 2 of Figure 1) has a definite downstream imbrication. Sample 0-4, collected 1.1 miles downstream, shows no pronounced imbrication either upstream or downstream, although there is a marked preferred orientation. Beyond this point, from sample 0-6 on (1.6 miles downstream from 0-4) the imbrication is upstream, as is usually expected. An exception is sample 0-11,

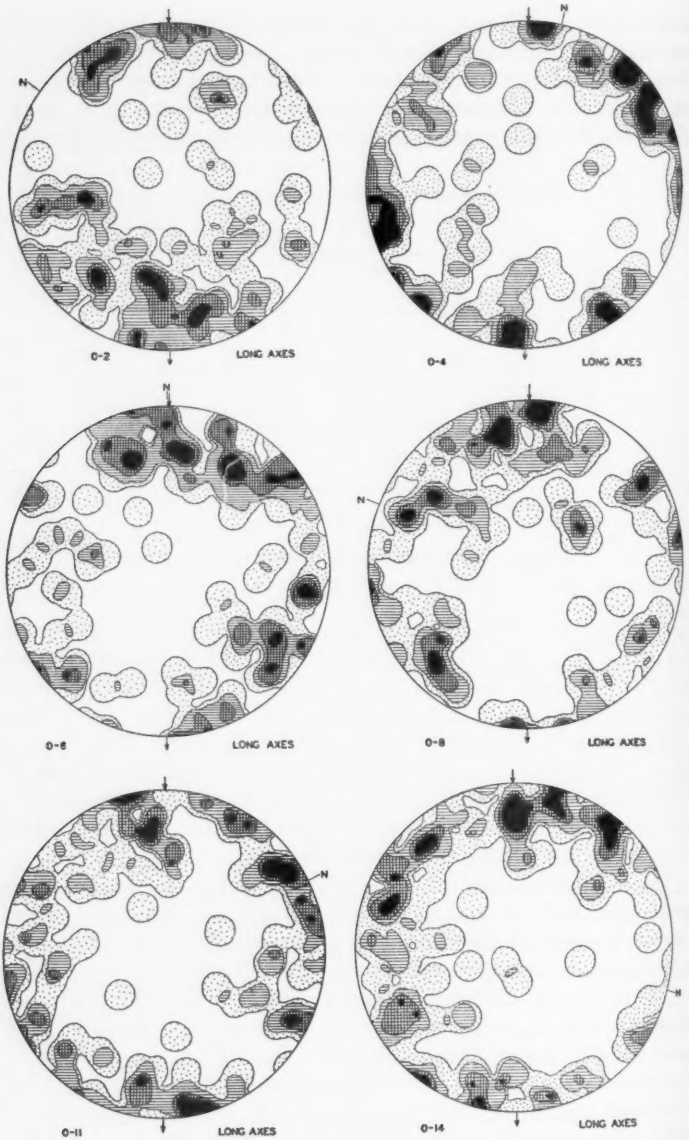

FIGURE 7.—*Petrofabric diagrams of San Gabriel gravel*
The sample numbers are shown in the lower left of each diagram.

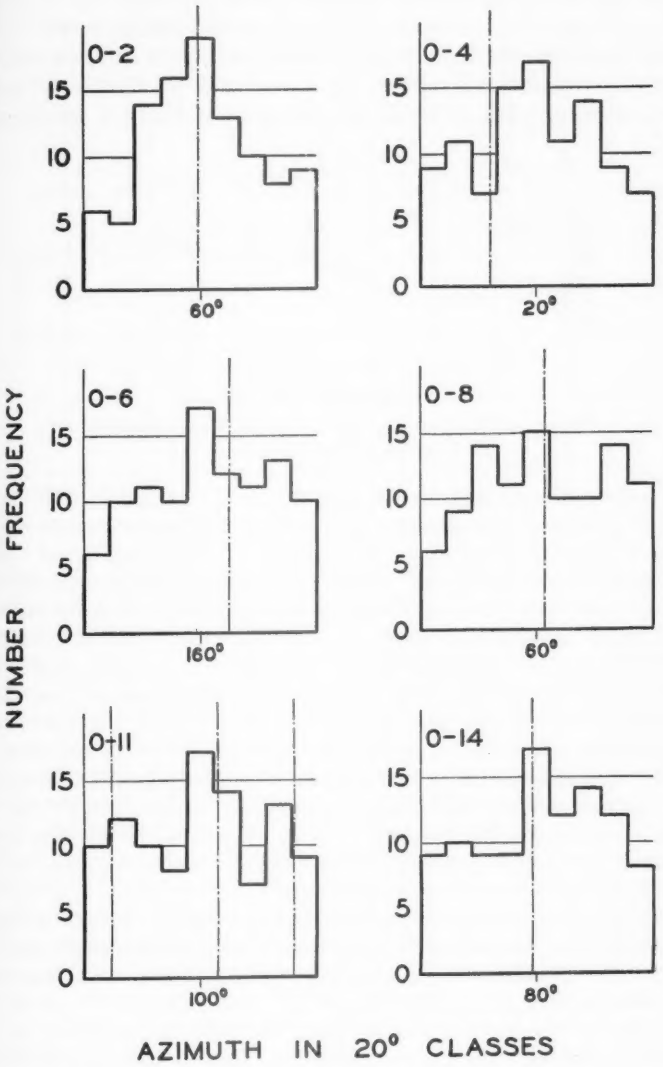

FIGURE 8.—*Azimuthal distributions of San Gabriel gravel*

The sample numbers are shown in the upper left of each histogram. Vertical broken line indicates valley trend.

which was an experimental sample and will be discussed. The apparent reversal of the imbrication has not been observed earlier, insofar as the writer is aware. The data do not represent an error in compass reading; they may represent current reversal in the lee of large boulders. Additional data regarding the imbrication are given in Figure 8, which shows

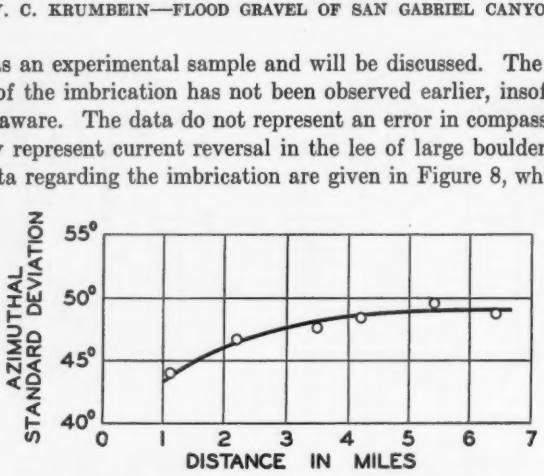

FIGURE 9.—*Standard deviation of azimuth as a function of distance along the valley*

the azimuthal distributions of the pebble axes arranged in 20° classes. The vertical broken lines indicate the position of valley trends with respect to the azimuthal distributions. The azimuth of the central modal class is given below each histogram. Sample 0-2 indicates that the modal class agrees well with the valley trend, whereas in sample 0-4 the trend is noticeably different from the mode. Such divergences indicate that it is not always safe to assume that the valley trend indicates the exact direction of current. Swirls, eddies, and even reverse currents are possible in floods, but Figure 8 indicates that statistically the valley trend does agree with current direction. This is strikingly brought out by sample 0-11, which was collected as an experiment from a sharp bend of the valley. The sample itself was collected from a bar-like feature trending N. 64° W., whereas the two valley trends were N. 31° E. and N. 7° W. The agreement of the central mode with the bar and of the other two modes with the valley trends is striking if coincidental.

The computed moments of distributions as irregular as those of Figure 8 are subject to several errors which depend upon the particular choice of class intervals used in the computations. A regrouping of the same data into classes with midpoints at 10°, 30°, 50°, instead of 0°, 20°, 40°, yields mean values as much as 5 or 6 degrees from the values found with the present grouping. The standard deviations, however, vary only one or two degrees. Usually the second moment is more sensitive to grouping than the first, but distributions as irregular as the present may yield a fairly large and consistent second moment regardless of the choice of particular

classes. Such variations may be avoided by computing the moments from the original ungrouped data.

Figure 9 illustrates the standard deviation of the azimuthal distributions as a function of distance along the valley. The data appear to show

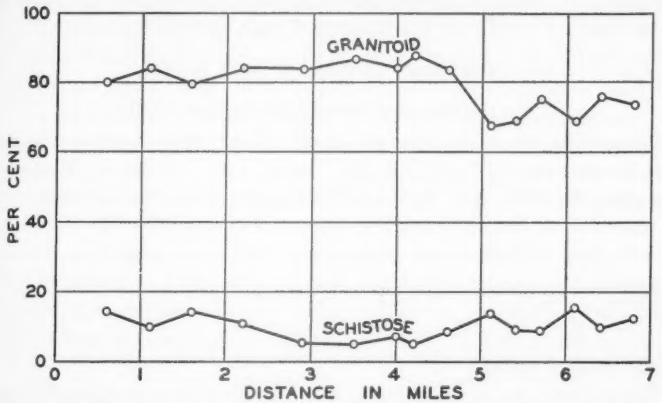

FIGURE 10.—*Variation in lithology as a function of distance along the valley*

a trend toward increasing values, which is indicated by the smooth curve. Geometrically the standard deviation of azimuth is a measure of the degree to which the pebbles are scattered about the preferred direction of orientation, and an increase in the value means an increase in the degree of spread. On this basis it appears that the extent to which the pebbles are oriented about their mean decreases downstream. This suggests that the control of orientation by the stream decreases as a function of distance.

Figure 10 shows the variation of pebble lithology as a function of distance downstream. The upper line represents the granitoid pebbles, which remain essentially constant over the first several miles but show a drop of about 13 per cent at the junction with West Fork. This suggests a contribution from West Fork of different percentage composition. The corresponding increase in the schistose pebbles (lower curve) is not enough to account for the change, but, when the dioritic pebbles and miscellaneous metamorphic and felsitic pebbles (Table 4) are included, the drop in the granitoid pebbles is balanced. The exact contribution from West Fork cannot be evaluated from the data at hand, but even if it were large the bulk of the pebbles must have been granitic. The junction with Bichota Canyon (Fig. 1) is not reflected in the pebble curves, which supports the field interpretation that the bulk of the material along North Fork came from Sycamore Flat or beyond.

The nature of the lithology curve has a bearing on the roundness data. The greatest amount of rounding occurred in the first several miles of North Fork, where the proportion of granitic pebbles is essentially constant. This supports the conclusion that the observed increase in roundness actually occurred during transit along the main valley, rather than by the influx of well-rounded material from a tributary.

INTERPRETATION OF THE DATA

SYSTEMATIC AND NONSYSTEMATIC VARIATIONS

A consideration of the data presented in the preceding section indicates that the characteristics of the deposits fall into two broad groups. The first group includes those features which appear to vary nonsystematically as functions of distance, and the second group includes those which appear to show definite trends along the valley. The mean size, the degree of sorting, the average sphericity, and the proportion of granitoid pebbles along North Fork are in the first group, whereas the average roundness, the imbrication, and the degree of preferred (azimuthal) orientation are in the second group.

The nonsystematic variation of the phi mean was mentioned earlier; if the laboratory data are reliable, there has been no systematic change in selective sorting of size along the valley, with the possible exception of the last samples in the series. This conclusion, it should be emphasized, refers to the intermediate sedimentation unit only. The graph of the phi standard deviation shows that the degree of spread of the data (the number of grades in the sediment) remains fairly constant along the valley, including even the last samples in the series. It should be noted that two apparently independent variables are included in the concept of selectivity with respect to size. Not only may there be a selection (or lack of selection) on the basis of size itself, but also on the basis of how closely grouped the particles are with respect to the mean size. It is interesting to note in connection with the phi mean and the phi standard deviation that the standard deviations of the 15 values about their own means are such that 10 observations fall within one standard deviation of the means, and 5 fall without. This is roughly the proportion one may expect in nonsystematic fluctuations.

The average sphericity curve is somewhat erratic in its behavior but suggests a slight increase downstream. The observed increase is so slight, however, that it seems safer to include the data in the nonsystematic group pending further observations. Both the size and shape are related to settling velocity, so that the observed sphericity fluctuations are not inconsistent with the random variations in size. Strict parallelism is absent,

however, inasmuch as the sphericity data are confined to pebbles, whereas the size data cover the entire range.

The proportion of granitoid pebbles in the deposits shows a distinct decrease at and beyond the junction with West Fork, but within each section of the valley the variation is essentially nonsystematic. The abrupt change in lithology can be related logically to contributions from West Fork, so that the absence of a gradual trend in lithology, which would result from the denser pebbles being left behind, seems consistent with the other two characteristics which depend directly upon settling velocity. In this connection there may be a slight trend in the proportion of the schistose pebbles, as shown by the lower curve in Figure 10, but this trend is not sufficient to produce a marked effect on the proportion of granitoid pebbles.

RECONCILIATION OF SIZE AND ROUNDNESS DATA

There seems to be little question that the behavior of a particle during transport is controlled by its size, shape, and density; these three factors control the settling velocity of the particle. Hence it may be expected that if the velocity of the transporting medium shows a systematic variation along the valley, this change should be reflected by corresponding variations in the size, shape, and density of the deposited material.

The San Gabriel data suggest that no marked trends occur in size, shape, or lithology over the distance sampled, except for lithological changes accounted for by material brought in from West Fork. If the data and the reasoning are sound, depositional conditions must have been essentially constant along San Gabriel valley during some stages of the flood. This seems unreasonable at first glance, but it will be recalled that the samples were collected from a single sedimentation unit within successive log jams. As soon as a given jam was formed the velocity for a short distance upstream suffered an abrupt change, with deposition as an immediate response. Hence, if the successive log jams represent essentially the same dynamic conditions, the response of the load with respect to size, shape, and density would be essentially constant, with such nonsystematic fluctuations as the vagaries of chance introduced.

This response of the load to the dynamics attendant upon the log jams would not be affected by the roundness values of the pebbles at the point of deposition. That is, settling velocity is essentially independent of roundness, so that no particular selectivity of pebbles on the basis of roundness would occur behind the log jams. Instead, the selection based on size and shape would carry with it that particular average roundness which the pebbles have at the point of deposition. Inasmuch as the roundness is distinctly a function of distance (Wentworth, 1919; 1922) the aver-

age roundness at any point along the valley should reflect the total distance travelled, essentially independently of the exact conditions of deposition at the point where the pebbles were deposited. Size and shape, on the other hand, should reflect the immediate conditions of deposition.

Fortunately there is some independent evidence in support of these inferences. Dr. L. G. Straub informed the writer that the size response of sediments in experimental flumes often shows unexpected variations due to the immediate response of the load to velocity conditions. Indeed, it was this suggestion which offered the clue to the present analysis. Work by Marshall (1927) also has a bearing on the problem. A load of natural beach gravel was rotated in a drum for a distance equivalent to 320 miles, during which the weight of the material decreased 21 per cent due to abrasion losses, and yet the mean size and degree of sorting remained essentially constant. Hence considerable abrasion (and rounding) may take place in the absence of selective transport without affecting the size parameters of the load. The explanation apparently is that, as size reduction occurs, the material drops to the next smaller grade in such a manner that the geometrical shape of the size distribution remains essentially unchanged except at the extremes. The center of gravity and even the standard deviation may thus not be affected to any marked degree for a considerable distance.

If the load may thus remain fixed, any changes in size or shape observed along streams must be due to selective transportation in large part, rather than to abrasion, in the absence of "contamination" or "dilution" from tributaries. If selective transportation is absent (or rather if the selective factors remain fixed) the size and shape parameters may show merely nonsystematic changes en route.

This idea of the roles of selective transportation versus abrasion in the characteristics of deposits is not new. It has at least been touched upon by such writers as Pettijohn and Ridge (1932), Marshall (1927), Krumbein (1937), and Russel (1939).

ORIENTATION (FABRIC) OF THE DEPOSITS

The imbrication data are less conveniently integrated with the other data. The petrofabric diagrams (Fig. 7) display a varying degree of monoclinic symmetry, with the *a*-fabric axis roughly parrallel to the valley trend. The *a*-fabric axis is the axis of movement and should mark the direction of current at the point of deposition. The complete reversal of imbrication in sample 0-2 is difficult to explain if it represents an unmodified stream deposit. The stream gradient varies exponentially along the part of the valley sampled, as shown by a graphic test made from topographic map data (Krumbein and Pettijohn, 1938, p. 208-210)

so that no abrupt gradient changes occur, except those associated with log jams. The steep dip of the pebble axes and the downstream imbrication of sample 0-2 suggest post-flood slumping, but no evidences of disturbed bedding were noted. The sample was collected a short distance downstream from a large boulder, and it is conceivable that reverse currents in the lee of this obstacle may have controlled the imbrication. A similar study of other valleys may shed light on the question; at present it seems that the effect may be due to local causes.

A strong suggestion of systematic change in the degree of preferred orientation is present along the valley, however. Figure 9 shows that in general the degree of spread of the axes about their mean azimuthal trend increases downstream. The writer had expected the opposite effect because it seemed likely that as the pebbles become more rounded they could more easily be swung into parallel position by currents after deposition, so that on the whole the standard deviation should decrease. The observed effect is the opposite of this, however, and, if the data warrant generalization, it may mean that deposition was too rapid to allow later adjustments by the current, and the increased roundness permitted a more random scatter at the instant the pebbles rolled into position.

The reasoning here may be weakened by the effect of shape on imbrication. As roundness increases, the sphericity values usually increase, although the effect in the present instance is not marked. The degree of preferred orientation is in part a function of shape, at least in some deposits (Krumbein, 1939b), and the less spherical pebbles may be expected to show more marked orientation. On this basis the standard deviation of azimuth would normally increase downstream.

It is evident that additional data are needed before the present tentative reasoning can be accepted as establishing any generalizations. The observed data do support the contention, however, that the simultaneous study of several sedimentary characteristics as functions of distance affords a more complete albeit more complex picture than does the study of any single attribute.

THEORY OF PEBBLE ROUNDING

DEVELOPMENT OF AN EQUATION FOR ROUNDING

Among the attributes of the flood deposits described in the preceding section, the roundness data afford the clearest picture of a systematic change with distance, largely independent of the particular conditions at any given site of deposition. In this section an attempt will be made to develop an analytic theory of rounding as a function of distance, to account for the observed variation. This theory will be developed for

a limited size range (about 25 to 70 mm. diameter), but its implications in terms of other sizes will be touched upon.

The curve in Figure 6 suggests that pebble rounding is a rapidly increasing function at the beginning of the process and becomes progressively slower as distance increases. That is, the more angular a pebble

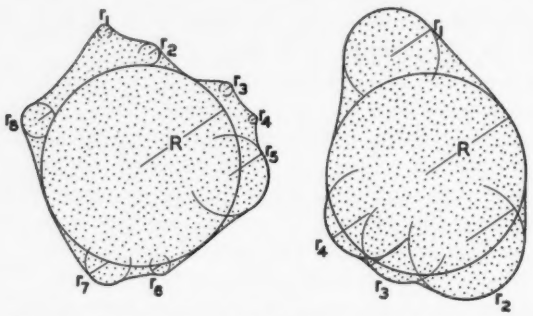

FIGURE 11.—*Diagram of pebbles, showing geometrical nature of roundness*

is, the more rapidly it is rounded. The continually decreasing slope of the curve also suggests that the pebble approaches some limiting value of roundness dependent upon the nature of the rock, the initial size and shape, and the rigor of the rounding process. The limiting value may thus be considered as an asymptote to the graph of roundness as a function of distance.

As a first approximation to the physical process of rounding, attention is called to Figure 11, which represents the projections of two pebbles on a plane. The roundness is a numerical measure of the extent to which the average radii of curvature of the corners and edges (r_1, r_2, . . . r_N) approach the radius of curvature of the inscribed circle R. The process of rounding involves an increase of the small radii and a consequent increase of the ratio $\Sigma(r_i^2/N)/R$ (equation 3), which defines roundness. On the average it is easier for a sharp corner, with radius of curvature 2 mm., to be abraded to yield a radius of 3 mm. than it is for an edge of radius 14 mm. to wear down to a radius of 15 mm. That is, the larger radii require more time or effort for corresponding increases than do small radii. This affords a physical basis for the inference that the more angular pebbles are rounded more rapidly and that the process of rounding becomes progressively less rapid.

The foregoing analysis may be expressed mathematically by making explicit the assumption that the rate of change of roundness as a function

of distance (dP/dx) is proportional to the difference between the roundness P at any point x along the stream and the limiting roundness value P_o:

$$\frac{dP}{dx} = k(P_o - P) \quad \ldots\ldots (4)$$

Here k is a constant of proportionality. The derivative is positive because under this assumption the roundness will increase. Equation (4) is a differential equation which readily yields a solution. Rearrange the terms to obtain

$$\frac{dP}{P_o - P} = k \cdot dx \quad \ldots\ldots (5)$$

The left hand side of this equation yields $-log_e\ (P_o - P)$ on integration (Peirce, 1929, formula 26), and the right hand side becomes kx. The solution of equation (5) therefore, is

$$-log_e\ (P_o - P) = kx + C \quad \ldots\ldots (6)$$

where C is the constant of integration. To evaluate the constant, assume[1] that the roundness P is zero at $x = 0$, so that

$$C = -\ log_e\ P_o.$$

The value for C is substituted in equation (6) to obtain

$$-\ log_e\ (P_o - P) = kx - log_e\ P_o.$$

Rearrange and change signs:

$$log_e\ (P_o - P) - log_e\ P_o = -\ kx$$

or,

$$\log_e\left[\frac{P_o - P}{P_o}\right] = -\ kx \quad \ldots\ldots (7)$$

The antilog of equation (7) is

$$\frac{P_o - P}{P_o} = e^{-xk}$$

which may be simplified to yield the standard form

$$P = P_o\,(1 - e^{-kx}) \quad \ldots\ldots (8)$$

Equation (8) represents the theory of rounding and states that the roundness of the pebble at any point along the stream is equal to the final roundness value times unity diminished by the exponential e^{-kx}, where k may be considered as the coefficient of rounding.

[1] The theory is not dependent upon this assumption for its validity, inasmuch as one may assume an initial roundness $Pi > 0$ and obtain an equivalent but more complicated function.

The observational data may be tested by means of equation (7), which yields a straight line when $(P_o - P)/P_o$ is plotted on a logarithmic scale against x on an arithmetic scale. For this purpose a value is assigned to P_o; for the San Gabriel gravel the value 0.450 was chosen because it

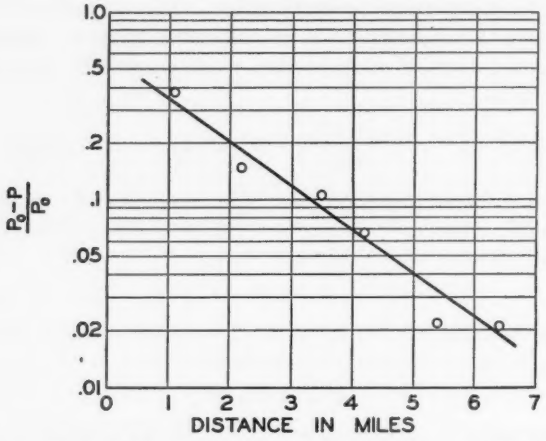

FIGURE 12.—*Graphic test of rounding theory with San Gabriel data*

is slightly larger than the largest observed value, 0.441. Figure 12 is the corresponding graph and shows a reasonably good agreement. The most notable departure is the fifth point (sample 0-11), which is due to the fact that that sample had a slightly larger value than sample 0-14 (0.441 against 0.438). The graphic test is very sensitive to values near the asymptote, and points slightly out of line are strongly affected.

To evaluate k, one chooses that value of x for which $(P_o - P)/P_o = \frac{1}{2}$. Call the x-value $x_{\frac{1}{2}}$. Then in equation (7),

$$\log_e (1/2) = -kx_{\frac{1}{2}}.$$

Change signs, and recall that $-\log_e (1/2) = +\log_e 2 = 0.693$. Hence

$$kx_{\frac{1}{2}} = -\log_e (1/2) = \log_e 2 = 0.693$$
$$k = (0.693)/x_{\frac{1}{2}} \quad \ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots (9)$$

The value of $x_{\frac{1}{2}}$ can be read approximately from Figure 12, although more detailed paper should be used. At $x = 1$, the ordinate is 0.34. Half this value is 0.17, which is found at $x = 2.4$. Hence the half distance is 1.4 miles. Substitute this for $x_{\frac{1}{2}}$ in equation (9) and obtain $k = (0.693)/(1.4) = 0.5$ approximately. The particular values of P_o

and k for the San Gabriel gravel of size 25-70 mm. diameter are, by substitution in equation (8):

$$P = 0.45(1 - e^{-0.5x}) \quad \ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots (10)$$

Physically this means that the limiting value of roundness is 0.45, and the coefficient of rounding is 0.5. It should be borne in mind that this

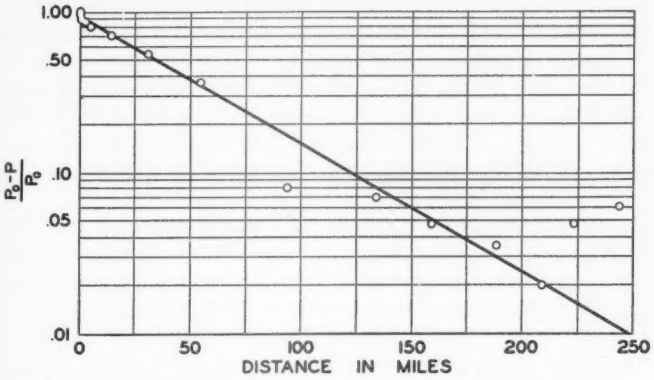

FIGURE 13.—*Graphic test of rounding theory with Wentworth's data*

analysis applies only to pebbles in the size range 25 to 70 mm. The values given in equation (10) do not apply to the flood deposits as a whole; if average roundness were measured for all sizes present in a bulk sample the numerical results would undoubtedly be much different. It is likely that a much lessened rate of rounding would be found. The present simplified analysis suggests that the two parameters P_o and k may themselves vary as functions of size and other factors.

The theory of rounding as developed here was recognized at least qualitatively a number of years ago. Wentworth (1919) studied the abrasion of marble cubes in a tumbling barrel and later (1922) studied quartzite pebbles along a stream. In both cases curves similar to the present were obtained, although the field data were complicated by the size factor. This limitation did not affect the marble cube study, however. Fortunately Wentworth's definition of roundness is enough like Wadell's in its geometrical significance that the earlier data may be used as a further test of the analytical function. One exception must be noted, however. Wentworth's data showed a decrease in roundness after a maximum was obtained. The present theory does not consider this possibility so that for an asymptote the value 0.88 was chosen, and a graphic test performed on the rising part of the curve. Figure 13 shows

the result, carried two points beyond the maximum in Wentworth's curve, to show the wide departure from theory as the original data drop below the maximum value beyond $x = 200$ miles. The agreement is satisfactory except for a single value which falls noticeably below the straight line. In the writer's opinion this is an exceptional reading; it is surprising that the data agree so well on the whole, when it is recalled that Wentworth's measurements involve only the most convex corner of each pebble, and the averages are based on four pebbles. A repetition of the experiment with a larger load may be expected to yield a closer agreement. Wentworth's later field study yields a fair agreement with theory also, except that the first few observations depart widely from the straight line. A further study of the factors in pebble rounding is contemplated, and discussion of initial departure from theory will be deferred until it is completed.

It is interesting to note that Wentworth's marble cube data yield a value of $k = 0.02$ and $P_o = 0.88$. In comparison with the San Gabriel granite pebbles this implies that the marble cubes reached a much higher limiting value (0.88 against 0.45), but at a much reduced increment of rounding (0.02 against 0.5). There is a fertile field for research in the work begun by Wentworth in 1919, and his original paper offers many sound suggestions which present workers will find profitable.

IMPLICATIONS OF THE THEORY

A logical extension of the present theory is a study of the roundness functions of other size pebbles and particles. A qualitative examination of the several grades in the 15 channel samples shows that the rate of rounding is definitely a function of size, and that the smaller grades show decreased rates of rounding; the sand sizes showed no appreciable change. There are two parameters involved: the coefficient of rounding k, which controls the rapidity of the process, and the limiting value, P_o, which controls the end product. Either or both of these may change with original size, shape, rigor, and rock type. The problem is directly amenable to quantitative attack, and from a study of each variable as it affects rounding, much light will be shed on the characteristics of sediments which may be extremely valuable in the study of ancient deposits. A further extension of rounding theory involves critical velocities beyond which particles may be broken by impact, so that roundness actually diminishes in the direction of transport (Russell, 1939). Complexities are likewise introduced by possible selective factors in transportation which may result in low-sphericity particles outstripping high-sphericity ones in suspension transport, and the reverse in traction transport. Relative roundness for a given shape may be a factor here also (MacCarthy, 1933; Russell, 1939).

There is an additional complexity in the field evaluation of P_o and k for any given size. During rounding the particles are reduced in size, and, if the study is made in terms of a single grade as in the present instance, the end-product pebbles may represent larger particles worn down to that grade size during transport. Such factors as this must perhaps be evaluated in tumbling barrel experiments; the agreement of both the San Gabriel data, where the factor may have been present, and Wentworth's data, where the factor was present, suggests that the essential nature of the theory is not involved, although it may mean that values of k and P_o will be only approximate in field studies.

SUMMARY AND CONCLUSIONS

The present study is an attempt to isolate some of the factors in flood environments which may reflect themselves in the characteristics of the deposited sediments. Work was confined to a single sedimentation unit which could be conveniently correlated from sample to sample. The findings therefore are confined to part of the sequence of events only. It is not possible to pass final judgment on the variables involved, but it does appear that some factors vary in a non-systematic fashion along the stream whereas others show systematic changes as functions of distance. Size and sorting within a single sedimentation unit may be subject to nonsystematic variations, unless the sampling or laboratory errors were so large as to cloud this point. Whether a size gradient would be disclosed if the entire thickness of deposits were analyzed is not known, although from the field examination it appeared that at least the upper depositional unit showed a progressive size decrease downstream. Roundness, on the other hand, appears definitely to show a gradient, independent of precise conditions of deposition along the valley.

Other problems raised by the study are whether mean size and degree of size sorting are independent of each other; whether imbrication undergoes a systematic variation downstream; the exact nature of the functional relation between rate of rounding and mean size; and the problem of the undoubtedly complex roundness relations in a mixture of sand and gravel, as in the present deposits. These problems may be attacked in the field and also experimentally; the writer plans to extend his work to include both aspects of the problem. Problems of sampling are also emphasized much more strongly by modern studies and they must be solved if generalizations of natural phenomena are to be placed on a sound foundation.

The application of the methods described in this paper to ancient deposits brings with it additional problems, among which those of sampling are perhaps the most important. Dependence upon accidental

outcrops, inability to find traces of such features as log jams, and other factors render the complete evaluation of ancient conditions much more difficult. Reliance may be placed on more closely spaced samples, perhaps, but the statistical problems will have to be worked out before definite progress can be achieved. The present study does suggest that roundness may furnish important data, and it may be expected that systematic changes in selective factors, dependent upon gradual velocity changes rather than abrupt effects as in floods, may result in definite size gradients in normal fluvial deposits.

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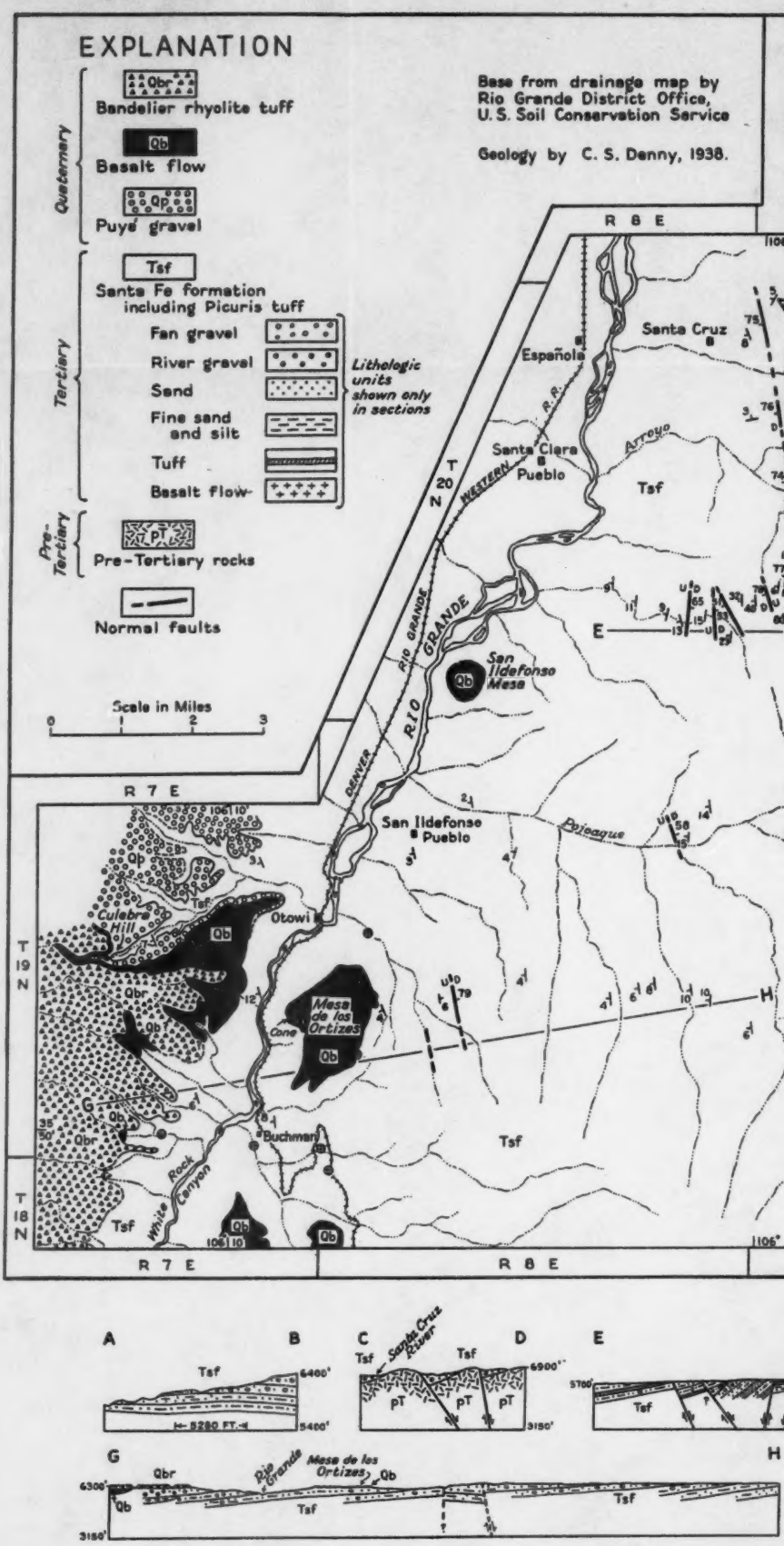

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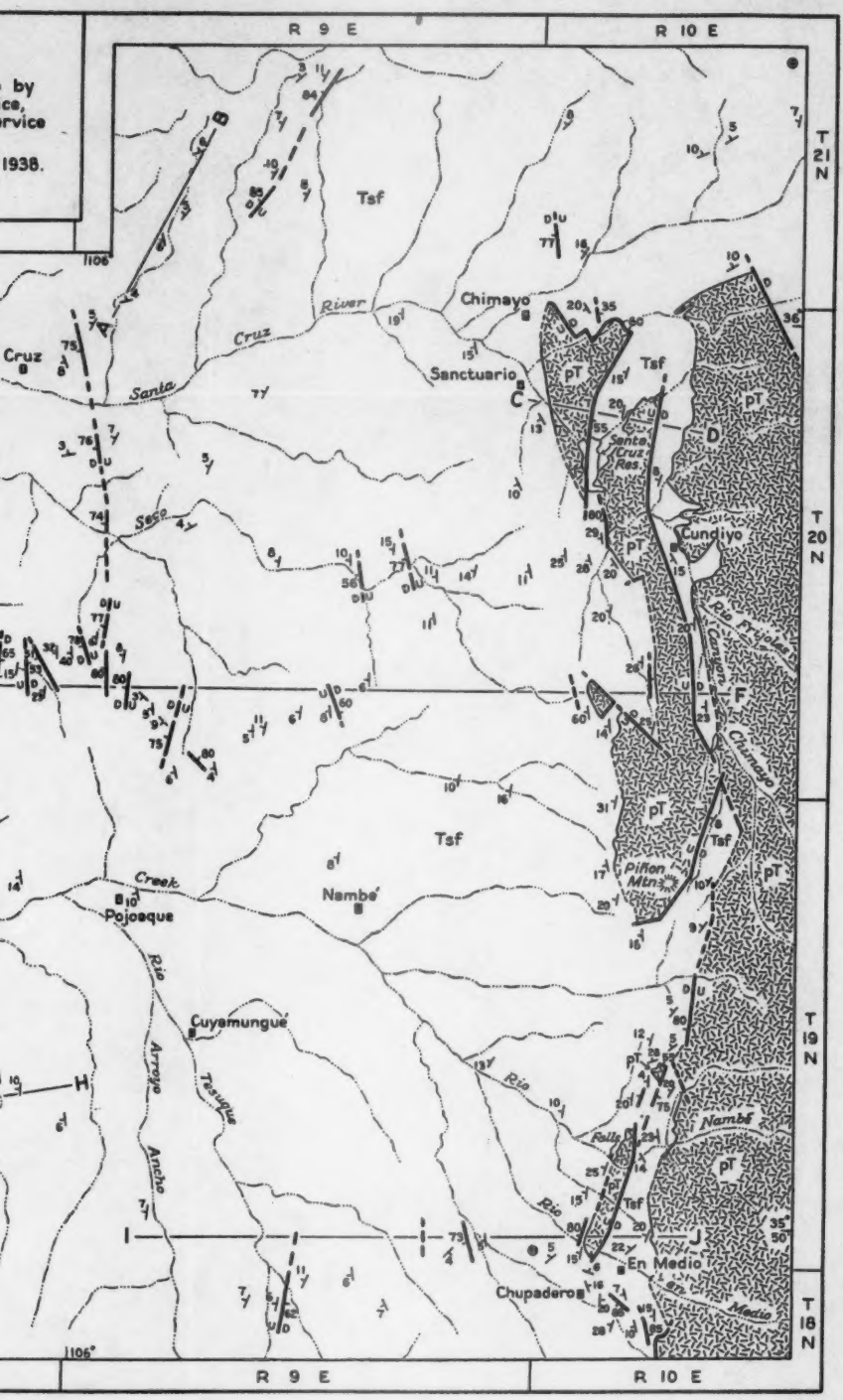

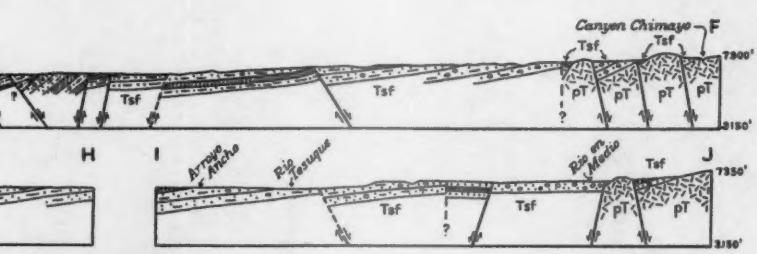

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LF OF ESPAÑOLA VALLEY, NEW MEXICO

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SANTA FE FORMATION IN THE ESPAÑOLA VALLEY, NEW MEXICO

BY CHARLES S. DENNY

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ABSTRACT

The type area of the Santa Fe formation (late Tertiary) is the region north of Santa Fe, New Mexico, between the Sangre de Cristo and Jemez mountains. The formation includes gravel, sand, silt, volcanic ash, and a little clay. Intraformational breccias are abundant in certain zones. Most of these materials were laid down as coalescing alluvial fans in a basin which, at its maximum, extended to the east and probably far to the west of the present limits of the formation. This alluvium was deposited by streams which flowed westward from highlands that bordered the basin on the east. Throughout most of the period of deposition these highlands were relatively low. During late Santa Fe time the eastern highlands stood at a somewhat greater elevation relative to the basin floor, and river gravels indicate that the center of the basin was occupied by a through-flowing river. The events of post-Santa Fe time have either removed or concealed all trace of the deposits laid down in the western part of the basin. Post-Santa Fe deformation has given the beds a general westward dip at angles of 5 to 10 degrees or more. At many localities the beds are displaced by normal strike faults, many of which are filled with conglomerate or sand dikes. Similar faults displace the unconformity which separates the formation from the older rocks to the east. Along their west border the Santa Fe beds are covered by the younger sediments and volcanics of the Jemez Mountains.

INTRODUCTION

The unconsolidated alluvial deposits which partly fill the Rio Grande Depression near Santa Fe, New Mexico (Fig. 1), were first described by Hayden in 1869 and were called the "Santa Fe marl". Fossils found by Cope (1874, 1884) were assigned to the middle Miocene epoch. An Upper Miocene and Lower Pliocene age is recorded by later workers (Matthew, 1900, p. 19-75; Osborn, 1909, p. 65; 1918, p. 34; Simpson, 1925, p. 230; Frick, 1933, p. 505-652). In his latest paper Frick (1937, p. 6) lists all the fossil localities under the heading "Late Tertiary" or "Middle Miocene to Upper Pliocene". Although early workers interpreted the formation as a lacustrine deposit, recent studies support the fluviatile hypothesis originally proposed by Davis (1900) and elaborated by Johnson (1903). Throughout northern New Mexico most of the unconsolidated deposits of the Rio Grande Depression were originally considered to be one formation. For example, on the geologic map of New Mexico by Darton (U. S. Geol. Survey, 1928) these deposits are grouped together as one formation assigned to the Miocene and Pliocene. However, recent studies have shown that this unconsolidated alluvium is divisible into several formations, some of which lie above and some below the fossil-bearing beds of the type region. In the Abiquiu quadrangle Smith (1938) found two formations, the Abiquiu tuff and the El Rito formation, which are stratigraphically below the beds (Santa Fe formation) which he correlates with Hayden's "Santa Fe marl". Cabot (1938)

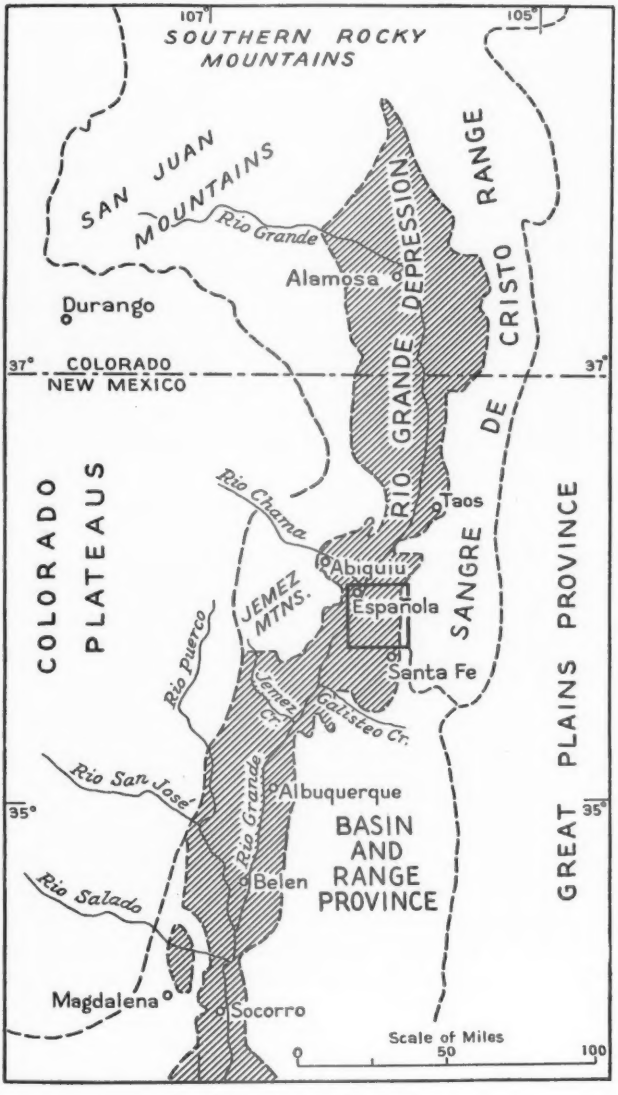

FIGURE 1.—*Index map*

Showing northern part of Rio Grande Depression and area discussed.

found a tuffaceous deposit, the Picuris tuff, below the "Santa Fe marl" in the Picuris re-entrant (Fig. 1. East side of Rio Grande, east of Abiquiu).

The name, Santa Fe formation, is now used for those deformed and predominantly fluviatile deposits of the Santa Fe region that contain the vertebrate fauna. Similar deposits which contain a few vertebrate remains are found throughout most of the Rio Grande Depression in New Mexico and are correlated with the type region (Bryan, 1938, p. 205). These deposits are probably the equivalent of the Gila conglomerate of southern Arizona (Gilbert, 1875; Knechtel, 1936).

Although extensive collections of fossils from the Santa Fe formation have been gathered in the region between Santa Fe and Taos, except for the study of the Abiquiu area by Smith (Smith, 1935, 1938) and that of the eastern contact of the formation by Cabot (1938), no detailed map of the formation in this area has been published. The writer devoted about six weeks to a study of the formation in the southern half of the Española Valley and compiled the geologic map (Pl. 1) from data recorded on aerial photographs and drainage maps of the U. S. Soil Conservation Service. The limited time available for field work, the variability of the formation, and the abundance of faults make it impossible to present a complete stratigraphic column for the Santa Fe formation in the type region or to give a precise figure for the total thickness of rocks exposed. The various subdivisions of the alluvial fan deposits cannot be mapped as distinct lithologic units and are given below for the convenience of description.

ACKNOWLEDGMENTS

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LITHOLOGY

ALLUVIAL FAN DEPOSITS

Gravel and coarse-grained sand.—The gravel is irregularly stratified, and most outcrops show cross-bedding (Pl. 2, Fig. 1). Individual strata are from a few inches to 10 feet or more in thickness, and in most places

FIGURE 1. CROSS-BEDDED SAND AND GRAVEL
Exposure in arroyo bank about 2 miles northeast of Nambé.

FIGURE 2. IRREGULAR LENSES OF GRAVEL IN SAND
Edge of mesa about 2½ miles north of Chimayo. (Shadows at base of gravel lenses).

COARSE-GRAINED FAN DEPOSITS

FIGURE 2. INTRAFORMATIONAL BRECCIA
Angular and rounded fragments of clay in stratified sand.
Locality same as Figure 1.

FIGURE 1. SAND AND SILT WITH INTRAFORMATIONAL BRECCIA
South side of Arroyo Seco about 4 miles east of Rio Grande.
Breccia above hammer.

FINE-GRAINED FAN DEPOSITS AND INTRAFORMATIONAL BRECCIA

the coarse-grained members are the thickest. The constituent pebbles, cobbles, and boulders are angular and subangular. Colors vary from light gray to reddish-brown. The stones in the gravel are predominantly granitic. Limestone, sandstone, and volcanic rocks occur in relatively minor amounts. Most of the gravel is partly cemented by calcium carbonate.

Bouldery gravel occurs only near the eastern limits of the formation. In an arroyo about 1½ miles north of Nambé Falls (Pl. 1; T. 19 N., R. 10 E.) this gravel rests unconformably on pre-Tertiary rocks. The boulders are angular and more than one foot in diameter. Only a small amount of matrix is present. A faint stratification can be seen. This deposit is perhaps a true fanglomerate (Lawson, 1913).

Lens-shaped masses of gravel interbedded with coarse and fine-grained sand are locally abundant. Exposures in cliffs show discontinuous beds of gravel which vary greatly in thickness and locally crosscut the stratification of the surrounding layers of sand (Pl. 2, Fig. 2).

The coarse-grained sand is unsorted. The individual grains are angular and consist predominantly of quartz, feldspar, and rock fragments. Individual strata range from one inch to several feet in thickness. Cross-bedding is visible in many exposures. The texture changes rapidly from bed to bed or from place to place in any one bed (Pl. 2, Fig. 1). Pebbles and fine sand are scattered through the coarse-grained sand.

Fine-grained sand and silt.—The fine-grained sand and silt form even layers which vary in thickness from a fraction of an inch to 10 feet. The individual beds can be traced for considerable distances. Cross-bedding is present. In some places the foreset beds are as much as 15 feet in length and dip at angles of only a few degrees. Perhaps such strata are eolian deposits. Layers of grayish-white, clean, well-sorted sand form thick, massive beds which may also be of eolian origin. Fragments of petrified bone are abundant on the outcrops of such beds.

Colors vary from white to gray to yellow to reddish-brown. Most of the sand is poorly sorted, and some strata contain a few pebbles. Resistant sand concretions from an inch to one foot in diameter are found in grayish-white sand.

Massive beds of brownish, poorly-sorted sand are partially cemented by calcium carbonate. Layers of brownish sand contain irregular masses and lenses which are partly cemented by calcium carbonate. The remainder of the layer is unconsolidated. Such strata resemble the beds of caliche which mantle erosion surfaces in the vicinity (Bryan and McCann, 1938, p. 5-6) and may be due to the evaporation of ground water during deposition of the sand.

Fine-grained beds are well exposed in the south bank of Pojoaque Creek just east of the highway bridge at Pojoaque and in the vicinity of the main highway from Pojoaque to Santa Cruz. Beds of tuff and intraformational breccia are locally interstratified with these fine-grained rocks.

Clay.—Except at a very few localities clay is not found in the fan deposits. A gray, calcareous clay (?) interbedded with sand and silt was noted at a point about 2 miles north of Chimayo (T. 21 N., R. 10 E.). Clay balls are found in fine-grained sand and silt at many localities.

Tuff.—Layers of white to gray tuff are intercalated with the fine-grained fan deposits. The tuff beds range from a few inches to 6 feet in thickness and can be traced for a mile or two along the strike. Except at a few localities the beds are disconformable on the underlying strata. Layers of unstratified tuff are probably the result of ash showers contemporaneous with deposition of the alluvial fans. Stratification in the tuff suggests reworking by streams. Tuff beds are abundant in the vicinity of the main highway from Pojoaque to Santa Cruz.

Intraformational breccia.—At several localities intraformational breccia is found intercalated with fine-grained sediments. These breccias consist of fragments of vari-colored clay and silt scattered throughout medium-grained sand that shows definite stratification (Pl. 3, Fig. 2). The breccia fragments are angular, oval, round, or slab-like, and range in length from one-half inch to 6 feet. Most fragments are less than one foot in diameter. At one locality a cross-bedded sand contains balls of clay as much as one foot in diameter, and irregular blocks from 2 to 3 feet wide and 6 feet long. The breccias do not exceed 50 feet in length nor about 20 feet in width. They rest disconformably upon even-bedded, fine-grained sand or silt (Pl. 3, Fig. 1).

Many exposures show contemporaneous deformation of the strata which surround a lens of breccia. Cross-bedding and channel fillings of fine sand and silt, similar to those figured by McKee (1939, Pl. 2) from the lower part of the delta of the Colorado River, are locally associated with the breccias. Intraformational breccias are exposed in the south bank of Pojoaque Creek just to the east of the highway bridge at Pojoaque and also in a butte (Pl. 3, Figs. 1 and 2) on the south side of Arroyo Seco about one mile east of the highway crossing.

RIVER DEPOSITS

Gravel and sand.—The river gravel in the Santa Fe formation consists of stones which are either round, oval or sub-angular (Bryan, 1938, p. 207). A little brownish-gray poorly sorted sand makes up the matrix

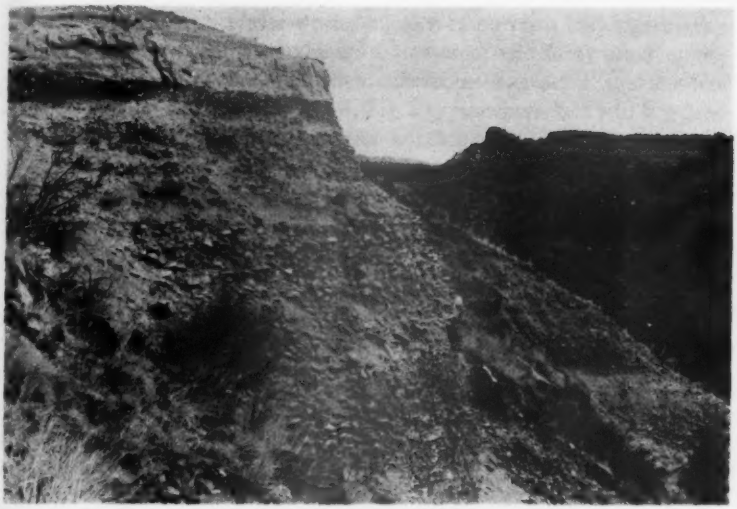

FIGURE 1. SANTA FE RIVER GRAVEL
North side of canyon 1½ miles west of Buchman. Largest boulders are one foot in diameter.

FIGURE 2. SAND DIKE ALONG FAULT PLANE
Prospect pit one mile east of Santa Cruz.

SANTA FE RIVER GRAVEL AND SAND DIKE

between the pebbles. Water-worn boulders of purplish and bluish-gray banded quartzite reach a maximum diameter of one foot; other stones reach a minimum of one-half inch. In some exposures the average diameter is from 6 to 8 inches (Pl. 4, Fig. 1); elsewhere it is 1 to 3 inches. Rock types include quartzite, quartz conglomerate, dark, fine-grained igneous rocks, a few pink "granites", fragments of tuff, and various volcanic porphyries largely of andesitic composition some of which are slightly water-worn. Individual strata range from 2 to 20 feet in thickness. A fine-grained sand is interbedded with the gravel. Cross-bedding was not observed. In some places the river gravel is interstratified with gravel of the alluvial fan type. In the area studied the only exposures of river gravel are in canyons on the west side of the Rio Grande (Pl. 4, Fig. 1) to west of Buchman (T. 19 N., R. 7 E.).

Silt and clay.—A greenish silty clay, which is associated with river gravel, was noted in one of the canyons to the west of Buchman. These outcrops were not studied.

VOLCANIC ROCKS

In this area volcanic rocks are almost lacking in the Santa Fe formation. Near Rio Nambé and Chupadero and to the west of Piñon Mountain (T. 18 and 19 N., R. 10 E.) there are outcrops of at least one basalt (?) flow, which ranges in thickness from 8 to 15 feet and is associated with tuffaceous gravel and sand. This sequence rests unconformably on pre-Tertiary rocks. These tuffaceous deposits lie conformably beneath typical arkosic Santa Fe sand and gravel, and perhaps represent a very early phase of Santa Fe deposition. They are described by Cabot (1938, p. 91) as a part of the Picuris formation.

On the west side of the Rio Grande near Buchman (T. 19 N., R. 7 E.) one or two flows of andesite-basalt are interbedded with river deposits. In the White Rock Canyon to the south of Buchman these flows increase greatly in number (Bryan, 1938, p. 207-208).

GEOLOGIC STRUCTURE

GENERAL STATEMENT

In northern New Mexico the Rio Grande Depression is a down-faulted basin (Fig. 2). That part of the depression which is known as the Española Valley (Bryan, 1938, p. 200) is bounded on the east and west by normal faults (Cabot, 1938; Smith, 1938). In this valley the Santa Fe formation dips westward, and the strata are broken by normal strike faults. Within the area studied (Pl. 1) the average strike of the beds is north and south, and the dip is westward at angles of from 2 to 10

degrees. North of the Santa Cruz River the strike is toward the northeast. Along the eastern contact of the formation most of the dips range from 10 to 35 degrees. No folds were observed.

The normal faults have an average dip of about 72 degrees. Conglomerate or sand dikes are to be found along many of these fault planes (Pl. 4, Fig. 2). The formation is in contact with the older rocks to the

TABLE 1—*Data on fault planes*

Description	Number of examples	Average dip	Percentage of total number of faults
Angle of dip from 0°–90°	49	72°	100
Angle of dip more than 60°	43	75°	87
Angle of dip more than 70°	33	79°	67
Santa Fe in contact with pre-Tertiary rocks	6	70°	12
Conglomerate or sand dikes	17	.....	35

east along a faulted unconformity. Such a structure gives rise to a series of westward-tilted fault blocks, bounded by normal faults (Structure sections C-D, E-F, and I-J; Pl. 1). To the west of the Rio Grande the formation disappears beneath the younger sedimentary and volcanic rocks of the Jemez Mountains.

FAULTS

The geologic map (Pl. 1) shows many eastward and westward-dipping normal faults which are only a small percentage of the total number. The abundance of faults in the structure sections and near the eastern limit of the formation, in contrast to the rest of the region shown on Plate 1, is the result of more detailed field work in those special localities rather than of any scarcity of faults elsewhere. The total displacement along most of the inter-Santa Fe faults is probably small. A summary of data on the faults is presented in Table 1. Conglomerate or sand dikes, from one to 10 feet wide, are found along many fault planes. Thin bands of silt with slickensided surfaces are locally associated with the sand dikes. Many dikes are composite. In some places there are fault zones which range from a few to 100 feet in width. These zones contain minor horsts and grabens, some of which are bounded by curved fault planes with dips along the individual plane varying from 40 to 90 degrees. Closely spaced parallel faults and small reverse faults are locally present. Mineralization is almost universally absent. A one-inch dike of calcite was noted along a minor break which has a total displacement of about 1½ feet.

RELATION TO SURROUNDING FORMATIONS

General statement.—In the southern part of the Española Valley the eastern contact between the Santa Fe formation and the pre-Tertiary rocks of the Sangre de Cristo Mountains is a faulted unconformity. The general nature of this contact from Santa Fe to Taos is described by Cabot (1938, Fig. 2). The average trend of the contact from Chupadero to Chimayo (Pl. 1) is north-south. The Santa Fe formation, which was laid down unconformably on the older rocks, has been tilted westward. The Tertiary beds are displaced along a series of normal faults which in some places parallel the unconformity and elsewhere cut across it at all angles. Later erosion of the Santa Fe beds has carved valleys which project into the foothills of the Sangre de Cristo Mountains to the east. The latter are underlain predominantly by granite and gneiss of pre-Tertiary age. North of Chimayo, outside of the area shown on Plate 1, the Santa Fe formation extends as a re-entrant into the pre-Tertiary rocks to the east ("Picuris re-entrant", Cabot, 1938, p. 98).

To the west of the Rio Grande the Santa Fe formation is overlain unconformably by a sequence of sedimentary and volcanic rocks of Quaternary age.

Nambé Falls-Chupadero section.—The Falls of the Rio Nambé (T. 19 N., R. 10 E.) are located in a small horst of pre-Tertiary rocks which is completely surrounded by the Santa Fe formation (Structure section I-J). To the east of the Falls there is an eastward-dipping (?) normal fault. To the west Tertiary gravel rests unconformably on the older rocks. The Santa Fe (Picuris tuff?) which is exposed to the east of the Falls can locally be seen resting unconformably on the pre-Tertiary rocks. At the southwest end of the horst, along the Rio en Medio, the Santa Fe rests unconformably on the older rocks. A few hundred feet to the west this unconformity is displaced along a westward-dipping normal fault (Structure section I-J).

A small outlier of pre-Tertiary rock is located about 1½ miles north of Nambé Falls. On the north and east sides of this outlier the Tertiary beds are down-faulted against it. On the west it is overlain unconformably by Tertiary gravel.

The eastern limit of the Santa Fe in the Nambé Falls-Chupadero section is an unconformity which is displaced along a normal fault at a point about a mile north of Rio Nambé and again about a mile south-southeast of En Medio.

Piñon Mountain section.—To the west of Piñon Mountain the pre-Tertiary rocks are overlain unconformably by the Santa Fe formation. In some places there are faults in the younger rocks close to the contact.

These breaks are not traceable for any distance along the strike and are therefore not shown on the geologic map (Pl. 1). A finger-shaped area of the Santa Fe formation extends northeastward from Piñon Mountain for a distance of about 1½ miles. The eastern boundary of this re-entrant in which the Tertiary strata are dipping westward was nowhere seen, but it is interpreted as an unconformity locally broken by faults. Piñon Mountain is bounded on the east by a normal (?) fault.

Another projection of the Santa Fe formation into the older crystalline rocks of the highlands is located about 2½ miles north of Piñon Mountain. This re-entrant is bounded on the east by an unconformity and on the southwest by a normal (?) fault (Structure section E-F). Near the eastern border of this re-entrant there is an easterly-dipping normal fault with a maximum displacement of about 20 feet (Structure section E-F). The similarity of a minor structure such as this with the major structure of the re-entrant is typical of the eastern contact of the Santa Fe formation throughout the area. About a mile to the east of this re-entrant there is a down-faulted wedge of late Tertiary sediment which extends southward from the village of Cundiyo. Both to the north and south of Cundiyo the Santa Fe formation can be seen resting unconformably on the pre-Tertiary rocks to the east. To the west of Cundiyo the Tertiary strata are cut off along a normal (?) fault.

Santa Cruz reservoir section.—The valley of the Santa Cruz reservoir (T. 20 N., R. 10 E.) is eroded in a down-faulted wedge of Santa Fe gravel. To the east Santa Fe beds rest unconformably on the pre-Tertiary rocks. The contact is clearly exposed in the arroyo which enters the northeast corner of the reservoir. To the south of the reservoir this unconformity is displaced along a westward-dipping normal fault which can be traced southward for a distance of about a mile where the contact is again an unconformity. To the west of the reservoir the Tertiary gravel is down-faulted against the older rocks (Structure section C-D) along the "Santa Cruz Reservoir Fault" (Cabot, 1938, Figs. 10 and 11). Uplift of the pre-Tertiary rocks along this fault has tilted westward the Santa Fe beds which rest unconformably on the older rocks to the east of Sanctuario.

Buchman-Otowi section.—On the western side of the Española Valley, to the west of Otowi and Buchman, the Santa Fe formation is overlain unconformably by sedimentary and volcanic rocks of the Jemez Mountains. These Quaternary rocks include the Puyé gravel, local basalt flows, and the Bandelier rhyolite tuff (Smith, 1935; Bryan, 1938). They were not studied in any detail, and their distribution on the geologic map (Pl. 1) is based on a brief reconnaissance and a study of aerial photographs.

THICKNESS

The total thickness of the Santa Fe formation in the type region cannot be accurately estimated. The strata are cut by many faults along which the total displacement is unknown. The structural interpretation which is presented in the structure sections (Pl. 1) calls for a section of Santa Fe which is at least 2000 feet thick in the center of the basin, whereas 3000 to 4000 feet is probably a closer estimate on the assumption that the displacements along the inter-Santa Fe faults are relatively small.

ORIGIN

AN ALLUVIAL FAN DEPOSIT

The Santa Fe formation, excluding the river-laid beds, is interpreted as an alluvial fan deposit (Bryan, 1938, p. 205). The distinctive characteristics of the formation are in summary as follows: There is a lack of sorting of both the coarse and fine material. The pebbles, cobbles, and boulders are angular and subangular. Most of the individual particles, whether large or small, are relatively unweathered. Rapid changes in texture, both in a horizontal and vertical direction, are the rule. The stratification of many of the coarse-grained members is indistinct. Cross-bedding and cut and filled channels are visible in many exposures.

Intraformational breccia is abundant in the finer-grained members of the formation. The origin of these breccias may be explained in the following manner. Thin layers of fine-grained sand and silt are deposited where the lower end of an alluvial fan merges with the central part of the depositional basin, which in this case is a river flood-plain. The even-bedding and widespread distribution of the fine-grained material suggests deposition in flood-plain ponds, rather than in broad stream channels. As the flow of water from the fan decreases, silt and clay are deposited. With desiccation mud cracks develop. The next stream of water from the fan picks up these fragments of silt and clay and cuts a shallow channel. The fragments are carried a short distance and deposited in a sandy matrix. The mud balls are fragments of mud which were rolled along the bottom of the stream. They are found in present-day arroyos such as the Rio Salado, Socorro County, and the Rio Chaco, San Juan County (Gardner, 1908; Bell, 1940). Perhaps some of the blocks now in a breccia originally fell from the undercut bank of a stream. The absence of foreign boulders, such as granitic rocks from the mountains, and of earthy material in the matrix indicate that the breccias are not mud flows (Blackwelder, 1928; Woodford, 1925). There is no evidence by which these breccias can be shown to be due to slumping on scarps developed along active faults.

The thick beds of grayish-white well-sorted sand that are found in association with the finer-grained members of the formation may be dune sand deposited near the lower end of an alluvial fan or on the edge of the flood-plain of a river (Smith, 1938, p. 956). Most of the gravel

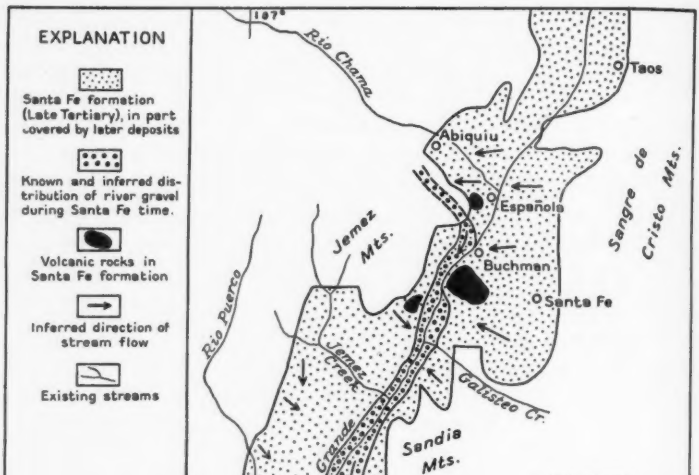

FIGURE 2.—*Map of Rio Grande Depression from Taos to Albuquerque*
Showing ancestral Rio Grande and inferred direction of stream flow during Santa Fe time. Modified from Bryan (1938), Fig. 46.

is partly cemented by calcium carbonate, and calcareous concretions are abundant in the finer-grained phases of the formation. These facts suggest deposition by streams rich in lime. The outcrops of limestone in the Sangre de Cristo Mountains to the east are the nearest available source.

The formation has many of the characteristics of the "Desert Environment", and the "Piedmont" and "Valley-flat" phases of the "Fluvial Environment" as given by Twenhofel (1932, p. 785-795; 800-811).

SOURCE OF SEDIMENT

The Sangre de Cristo Mountains to the east of the Española Valley (Fig. 2) are composed of rocks of two distinct ages. (*See* Darton's geologic map of New Mexico published by United States Geological Survey in 1928.) Granite, gneiss, schist, pegmatite, and similar rocks of

pre-Cambrian age (Darton, 1928, p. 272-274) are overlain unconformably by sedimentary rocks of Pennsylvanian age which include limestone and minor amounts of shale and sandstone.

The Jemez Mountains to the west of the Rio Grande are underlain predominantly by volcanic rocks of Cenozoic age. One of these, the Chicoma volcanic formation, includes "several thousand feet of andesite and latite lavas with some basalt, rhyolite and volcanic breccia" (Smith, 1938, p. 939). This formation is older than the Santa Fe and is probably "equivalent to the Miocene volcanics of the San Juan Mountains in Colorado" (Smith, 1938, p. 958). In most places the Chicoma volcanic formation is overlain unconformably by volcanic rocks of Quaternary age.

In the southern half of the Española Valley the Santa Fe formation contains an abundance of granitic gravel and arkosic sand. A few limestone pebbles are present. At most localities pebbles of volcanic rock are absent. These facts indicate that the Santa Fe formation was derived from the Sangre de Cristo Mountains to the east (Smith, 1938, p. 955-956). Near Chimayo (T. 21 N., R. 10 E.) arkosic Santa Fe gravel contains a few pebbles of volcanic rock. Such pebbles are not necessarily of western derivation (from the Jemez Mountains) because they could have come from outcrops of Picuris tuff (Cabot, 1938, p. 91) which are now either concealed beneath later deposits or have been removed by post-Picuris erosion. The slightly water-worn pebbles of andesite porphyry which are found in the Santa Fe river gravel to the west of Buchman were probably derived from the low-lying Chicoma volcanic center to the west.

AREAL DISTRIBUTION OF LITHOLOGIC TYPES

In general bouldery gravel is found only near the eastern limit of the formation from Chimayo to Chupadero (Pl. 1). As one goes westward from this contact the beds are finer and finer grained until one reaches about the longitude of Pojoaque and Cuyamungué. Northward from the above-mentioned villages runs a belt of rather fine-grained material which extends as far north as the Arroyo Seco where the belt turns slightly to the northeast (parallel to the strike of the rocks) and continues to the Santa Cruz River. To the west of this belt the formation is somewhat coarser-grained, although nowhere is it as coarse as near its eastern margin. Near the Rio Grande the Santa Fe is mostly sand and silt although layers of pebbly fan gravel can still be found interbedded with the Santa Fe river gravel to the west of Buchman.

ANCESTRAL RIO GRANDE

The gravel which crops out to the west of Buchman is interpreted as the deposit of a large, through-flowing river on the basis of the following

evidence (Bryan, 1938, p. 205-208): (1) The abundance of large water-worn fragments of resistant rock in contrast to the angular stones in the surrounding fan deposits; (2) the presence of greenish clay (flood-plain deposit); and (3) the geographic location of this gravel with respect to the bordering fan deposits of the Española Valley and to similar river gravel in adjoining areas to the south (Fig. 2). The well-rounded boulders of quartzite and volcanic porphyry, found near Buchman, are too greatly water-worn to have come from the surrounding fans of Santa Fe time and must have been carried by the late Tertiary river from extensive outcrops of quartzite and volcanic rock in northern New Mexico.

The hypothesis that the Santa Fe formation was laid down in several contiguous basins, the axis of which was marked by a through-flowing river, is based on field studies by Bryan and others (Bryan, 1938; Bryan and McCann, 1937; Denny, 1940; Smith, 1938). River gravel is found at the south end of White Rock Canyon, throughout the Santa Domingo Valley where good exposures of river gravel are found in road cuts along the Santa Fe-Bernalillo highway about 5 miles north of Algodones, and in scattered outcrops throughout the Albuquerque and Belen Valleys. The gravel shows a slight decrease in size from north to south.

The river gravel is locally associated with lava flows which appear to have confined the course of the river to some extent (Fig. 2).

The controlling base-level for the deposition of the Santa Fe formation in the Española Valley was the ancestral Rio Grande. This base-level was rising, probably as the result of tectonic disturbance at an undetermined point down river. In the southern part of the Española Valley the abundance of fine-grained material in the formation may be partly due to damming of the ancestral Rio Grande by lava flows, of which more than 1200 feet are exposed in White Rock Canyon to the south of Buchman (Bryan, 1938, p. 208).

ORIGINAL EXTENT OF THE FORMATION

Only the eastern part of the original basin of deposition of the Santa Fe formation is exposed in the southern half of the Española Valley. This fact is brought out in structure sections G-H and I-J because, as pointed out above, all the material, to the east of the outcrops of river gravel, was derived from the east. The formation must have originally extended some distance to the west of Buchman. Smith (1938, p. 956) states that there were low highlands "at the Chicoma volcanic center" during Santa Fe time. At its eastern edge the formation is not very coarse-grained except at a few localities. The rather gradual change in texture from the center of the original basin eastward to the present margin

suggests that the formation once extended a considerable distance farther to the east, perhaps nearly to the position of the present summits of the Sangre de Cristo Range. The small number of limestone pebbles in the Tertiary gravel, in spite of the relatively large areas of Pennsylvanian limestone to the west of the summit of the range, suggests that relatively early in the epoch of Santa Fe deposition these areas were more or less completely covered by detritus and, therefore, most of the Santa Fe gravel came from the pre-Cambrian rocks near the crest of the range. (*See* Darton's geologic map of New Mexico published by the United States Geological Survey in 1928. The areas of Magdalena limestone shown are somewhat exaggerated.)

CLIMATE OF SANTA FE TIME

The climate of Santa Fe time was probably semi-arid to humid with a rainfall somewhat greater than that of the present, which is 7 to 18 inches a year in the valleys (Cooperrider and Hendricks, 1937, Fig. 2). The great abundance of vertebrate remains attests to a more humid climate than that of the present as does the absence of saline deposits such as gypsum from the Tertiary beds. On the other hand the absence of carbonaceous matter in the finer-grained sediments, the presence of mud cracks, the unweathered character of most of the detritus, and the abundance of calcium carbonate suggest a semi-arid climate.

CONCLUSIONS

The Santa Fe formation contains alluvial fan and river gravel, sand, silt, clay, tuff, and intraformational breccia. The formation was deposited as coalescing alluvial fans which, during the late Tertiary, spread westward from the relatively low ancestral Sangre de Cristo Mountains toward the flood-plain of a southward-flowing permanent river, the ancestral Rio Grande. Gradients were low. To the west of the river there was a small highland area, the Chicoma volcanic center (ancestral Valles Mountains). Sand dunes were common along the eastern side of the flood-plain of the river. The climate was semi-arid to humid. Occasional showers of volcanic ash were spread over the surface, to be locally transported by rain wash and deposited near and on the river flood plain. Considerable vegetation was present in and near the flood-plain and also on the bordering highlands. An abundant vertebrate life inhabited the area, at least for part of the year.

Since the Pliocene the formation has been tilted westward and broken along many normal faults. Much material has been either eroded away or covered by later deposits. Extensive erosion has been in progress since at least the middle of the Pleistocene.

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GRANITES OF THE FRONT RANGE, COLORADO: THE INDIAN CREEK PLUTONS

BY MARGARET FULLER BOOS AND ESTHER ABERDEEN

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ABSTRACT

The Indian Creek plutons crop out around the northern and northeastern margin of the Rosalie lobe of the Pikes Peak batholith, west of Denver, Colorado. The granite is Silver Plume type and is the youngest pre-Cambrian intrusion of the Denver Mountain Parks region.

The granite appears to have come up along the edge of the mostly consolidated Pikes Peak mass and to have spread in a series of plutons and sills, the mode of intrusion being largely governed by the rock invaded. To the west and northwest, the magma spread a tongue of granite which wedged between quartz monzonite of the Mt. Evans region and its overlying schist. The magma flowed from the south and the southeast to the north and northwest. Erosion has since stripped off most of the schist above and has cut windows through the Indian Creek granite into the quartz monzonite below.

East of this wedge-like tongue, the granite worked its way up lit-par-lit into the Idaho Springs beds, boiled up locally in a series of stocklike centers or plutons, along concentric fracture zones, and spread between beds of schist and across them at low angles. The plutons and their major sills and dikes have a strikingly constant trend from northwest to southeast and lie in four sets, roughly parallel to the northeastern margin of the Rosalie lobe of the Pikes Peak batholith. Early egress of Indian Creek magma probably followed concentric cracks developed during the earlier intrusion of the Pikes Peak granite, and later intrusions followed fractures developed by the collapse of the Rosalie lobe.

When nearly crystallized, shear zones opened, northeast to southwest, in the Indian Creek plutons and permitted a little regenerated Pikes Peak magma to fill the fractures and to crystallize there, with its characteristic pasty textures.

INTRODUCTION

LOCATION OF THE AREA

The Indian Creek plutons and the interpluton and peripheral formations crop out on more than 200 square miles of the eastern slope of the Front Range, west of Denver, Colorado (Fig. 1). The topography is shown on the Denver Mountain Parks quadrangle of the United States Geological Survey, and a small part, at the south, on the Platte Canyon quadrangle.

FIELD WORK AND ACKNOWLEDGMENTS

The Indian Creek plutons constitute one of the six areas of younger pre-Cambrian granite in the Front Range (Boos, 1936). A reconnaissance field study was made in August 1935. In 1936, one of the 4 months devoted to mapping the younger pre-Cambrian batholiths of the Front Range (Boos, 1936) was spent on the Indian Creek plutons. For 10 weeks in 1937 and 1938, the structures of the plutons were detailed, and the geological map completed. The junior author was field assistant in 1936 and 1937, and Helen Dowling acted in that capacity in 1938.

The general geologic relationships, structures, mode of occurrence, and probable manner of emplacement of the plutons have been examined and are set forth in this paper. The geologic events which preceded, accompanied, and followed the intrusion of the plutons and adjacent crystalline rocks are summarized. Data were collected from more than 1500 field stations on the plutons and adjacent crystalline rocks. Rock analyses and examination of thin sections completed the study.

Field and laboratory studies were aided in 1936 and 1937 by grants from the Penrose Bequest of the Geological Society of America. Funds for the 1938 work came from the University of Denver. Sincere thanks are tendered these institutions for their generous assistance. To several geologists who reviewed the work, scanned the record and interpretations, and lent encouragement to the detailed study of one of the most complex areas of pre-Cambrian rocks in the Front Range, we acknowledge our debt with thanks.

GENERAL GEOLOGY OF THE AREA

GRANITIC ROCKS

Quartz monzonite-Mt. Evans batholith.—The oldest granitic rock is the quartz monzonite and related paragneiss of the Mt. Evans batholith. It is characteristically gray to blue gray, coarse, subporphyritic, granular, massive and locally gneissoid, and crops out to the west and northwest of the Indian Creek pluton area.

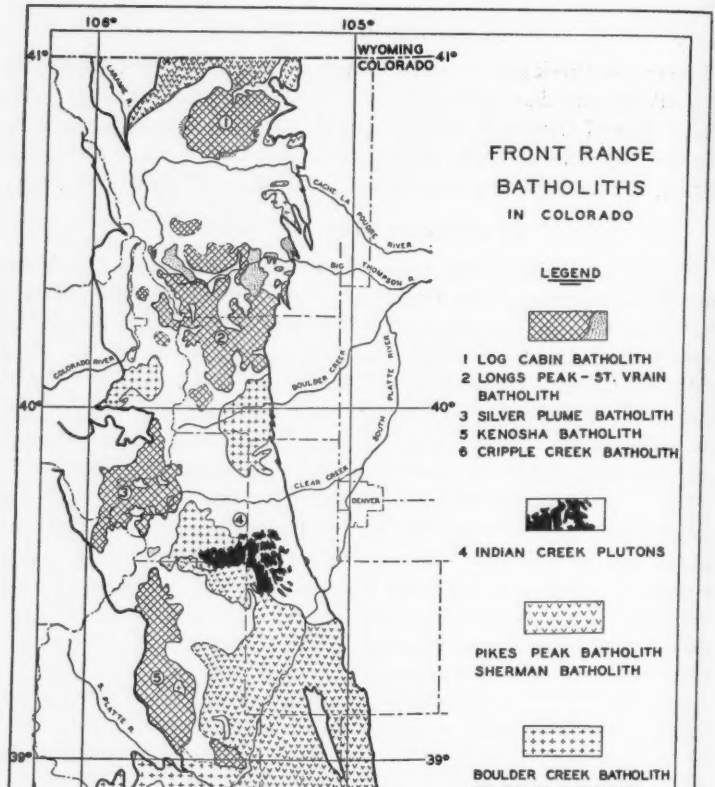


FIGURE 1.—*Front Range batholiths in Colorado*

Geology of the Log Cabin, Longs Peak-St. Vrain, Silver Plume, Kenosha, and Cripple Creek batholiths from maps of M. F. Boos, C. M. Boos, and E. Aberdeen. Other batholiths and intervening areas from field maps of M. F. Boos and adapted from geological map of Colorado (1935).

Pikes Peak granite-Rosalie lobe of Pikes Peak batholith.—This granite is younger than the quartz monzonite, which it intrudes, and generally older than the Indian Creek granite. The Pikes Peak granite is commonly spotty. Salmon pink to gray orthoclase and/or microcline, and gray to milky white quartz form a groundmass which is blotched with evenly spaced aggregates of black mica and hornblende. The groundmass is mostly even granular to dense, and gray, or brick red. Accessory apatite, magnetite, and sphene are common.

Indian Creek granite-Indian Creek plutons.[1]—The Indian Creek granite is, in field relationships, physical appearance, and crystalline character, closely related to the granite of the Silver Plume batholith 15 miles to the northwest. The rock is subtrachytoid to porphyritic, flesh to tan colored, and is distinguished by subparallel, similarly oriented lines of tabular idiomorphic feldspar crystals closely crowded and embayed by interstitial quartz, micas, and accessory minerals (Boos, 1936). Hornblende is absent. (See Table 1.)

The granite, named for Indian Creek Park where the rock was first identified (Boos, 1932), is the youngest pre-Cambrian granite in the area. It outcrops as a group of plutons—the Indian Creek plutons—which make four belts of subparallel slightly en echelon granite bodies concentrically aligned along the north and east margins of the Rosalie lobe of the Pikes Peak batholith (Pl. 1).

METAMORPHIC ROCKS

Idaho Springs formation.—The oldest rocks of the area are the metasedimentary schists and gneisses of the Idaho Springs formation which are

> "the groundwork into which the three successive and closely related granites described above made their way"

(Ball, p. 38, 1908). The formation was described for the Georgetown quadrangle as consisting of

> "interbedded metamorphic rocks divided into four intensely metamorphosed crystalline members equally metamorphosed and with a common structural relation to all other formations."

The description fits the schists and gneisses of Idaho Springs age exposed in the Indian Creek pluton region.

RELATIONSHIPS OF GRANITE AND METAMORPHIC ROCKS

The Idaho Springs beds were highly foliated before the invasion of the granites. The writers believe that the first magma—Mt. Evans

[1] Following the concept of Hans Cloos (1928), the term pluton is used to designate small bodies of deep-seated origin and is restricted, in this paper, to bodies of Indian Creek granite.

batholith—heated and softened the bulk of the Idaho Springs beds in the Indian Creek pluton area. To the north and northwest of this area, the Idaho Springs beds were intensively altered by the quartz monzonite of the Santa Fe Mountain area and the Mt. Evans batholith.

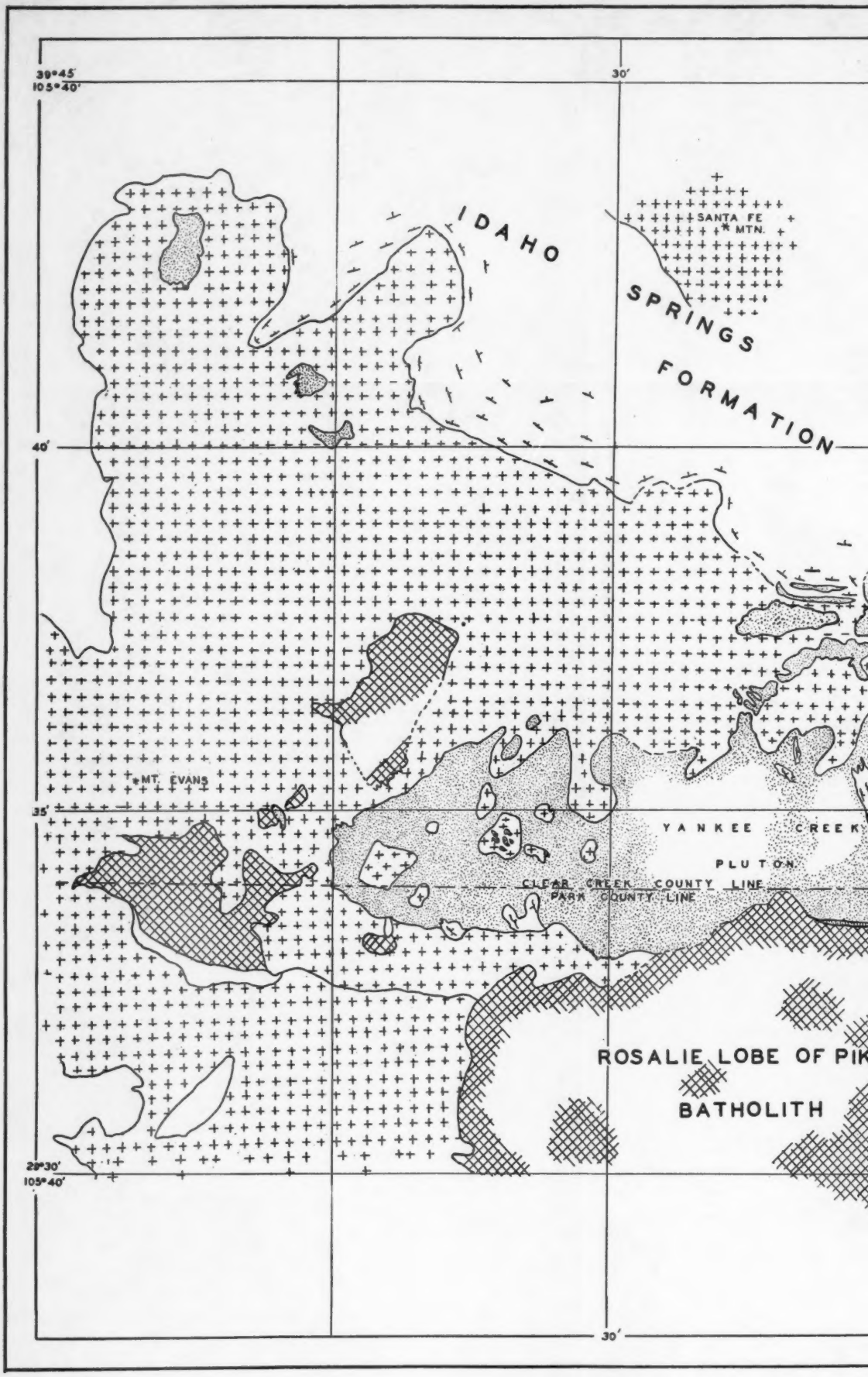
TABLE 1.—*Chemical analyses of the granites from younger pre-Cambrian intrusions of the Front Range*

	A	*B*	*C*
SiO_2	70.85	67.38	71.13
Al_2O_3	15.14	15.22	14.36
Fe_2O_3	.65	1.49	1.17
FeO	1.57	2.58	1.63
MgO	.64	1.12	.61
CaO	.71	2.12	1.03
Na_2O	2.44	2.73	3.14
K_2O	6.09	5.41	5.62
H_2O+	.77	.39	.42
H_2O-	.06	.01	.04
CO_2	.21	.18	.04
TiO_2	.31	.70	.25
ZrO_2	.....	.....	.02
P_2O_5	.30	.32	.18
SO_3	.....	.....	.....
Cl	.....	.....	.03
F	.....	.....	.....
S	.....	.06	.02
Cr_2O_3	.....	.....	.....
CuO	.....	.....	.....
MnO	.02	.04	.04
SrO	.....	.....	.....
BaO	.06	.14	.09
Li_2O	.....	.....	.....
Total	99.82	99.89	99.82

A. Longs Peak granite, Station 191 S. St. Vrain Highway. T. Kameda, analyst.
B. Silver Plume granite, Silver Plume. R. B. Ellestad, analyst.
C. Indian Creek granite, Buffalo Park, Denver Mountain Park. R. B. Ellestad, analyst.

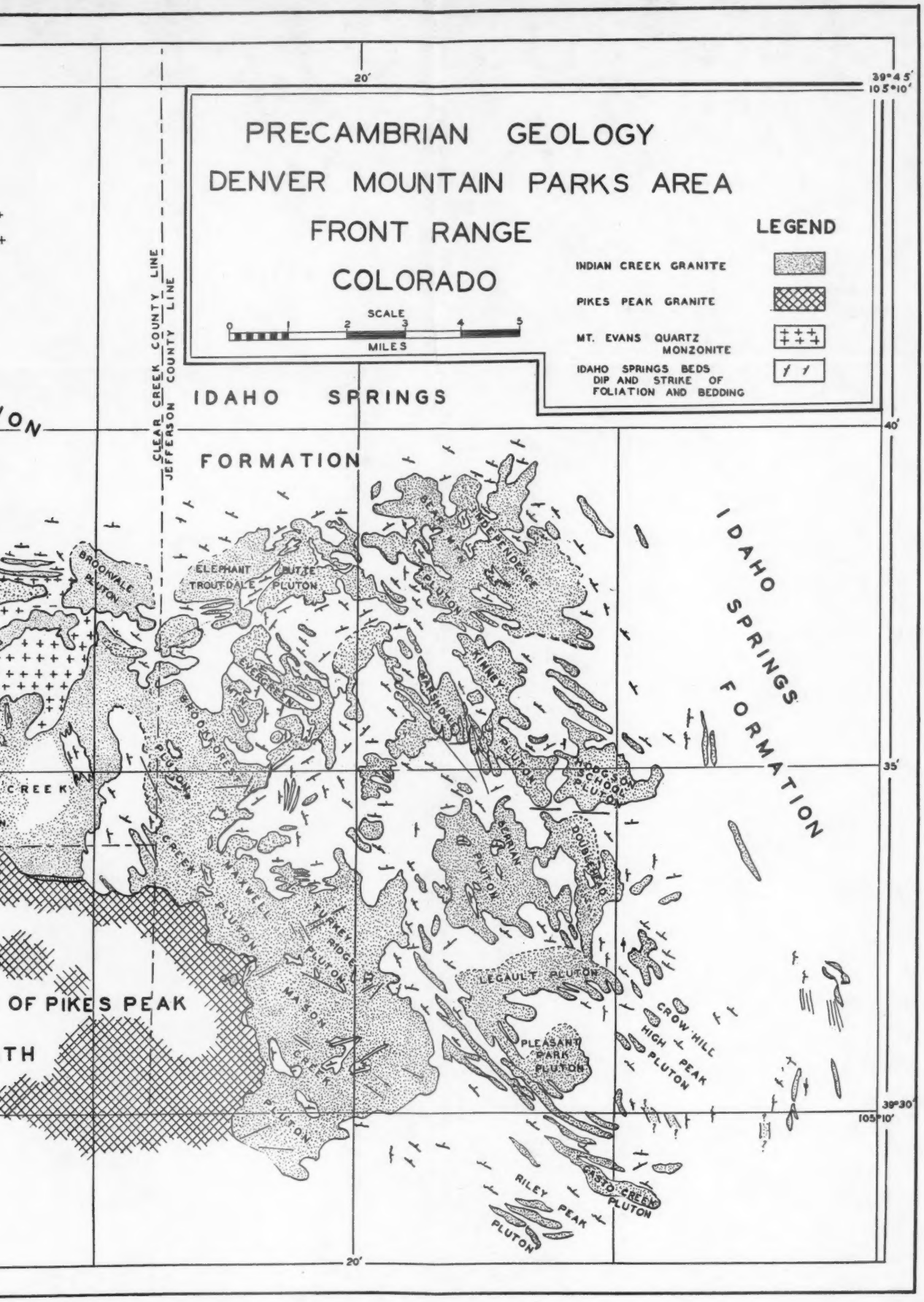
According to Ball (1908, p. 52), the monzonite magma stoped its way into the area, quarrying segments of Idaho Springs beds and engulfing the products.

The quartz monzonite crops out mostly to the west and northwest of the Indian Creek plutons. Locally, dikes and irregular bodies of Indian Creek granite transgress the quartz monzonite; one large lens-shaped pluton in the Yankee Creek area lies between the quartz monzonite and the overlying roof schists. Windows of the quartz monzonite underlying the lens have been exposed where Indian Creek and its tributaries cut through the granite pluton.



PRE-CAMBRIAN GEOLOGY OF DENVER MOUN

Geology by M. F. Boos, E. Aberdeen, and H. Dowling east of 105°30' W. Long. We



VER MOUNTAIN PARKS AREA, FRONT RANGE, COLORADO

. Long. West of this meridian the geology is modified from U. S. Geol. Survey, Prof. Paper 63, Pl. 2.

The second granite magma—Pikes Peak batholith—apparently did little to intensify contact metamorphism of the country rock in the Indian Creek pluton area. Few apophyses of either Pikes Peak granite or pegmatite extend into the surrounding formations. Most of the mineralising juices, retained within the batholith, had little direct effect on the country rock.

The third magma—Indian Creek plutons—was the great transformer and invader and wrought the most noticeable changes in the pre-existing rocks. It saturated the schists and gneisses adjacent to the Indian Creek plutons with its fluid. Aplitic and pegmatitic materials penetrated the Idaho Springs beds lit-par-lit and altered schists and gneisses 3 and 4 miles from the main plutons. Adjacent to the plutons the intense metamorphism obscured the records of earlier granitic invasions.

The Idaho Springs beds adjacent to the Indian Creek plutons have been converted locally into streaky migmatites which carry swirling "ghost lines", old drag folds, and older shear structures. Gneiss and schist are plastered over the summits of the granite bodies, enfold individual small plutons, and generally form the warp of pre-Cambrian fabric of the pluton area. The Idaho Springs beds encompass the area occupied by the Indian Creek plutons on the north and east and outcrop to the northeast for 10 miles beyond the pluton area. The foliation, close to the plutons and generally following the original bedding planes of the metasedimentary formations, dips away from the pluton area except at the south where it swings northeast. The regional trend of the Idaho Springs foliation and original bedding is mostly northwest—southeast and dips NE 20°-50°, 3 or 4 miles from the plutons.

Where the Indian Creek plutons and the Rosalie lobe of the Pikes Peak batholith adjoin, southwest of Black Mountain, there are many small subangular inclusions of the fine-grained, red, spotty Pikes Peak granite in the Indian Creek granite. A few well-defined apophyses of Indian Creek granite extend westward into the Pikes Peak granite selvage. Along Mason Creek, contact between the two granites is sharp, but, in most places, a distinct mica selvage 2 to 3 inches wide and a network of green veins on the Indian Creek side of the junction indicate that it is younger than the Pikes Peak granite (Anderson, 1924). Northwest and east of Black Mountain, the Indian Creek granite, thrust up over the Pikes Peak granite margin along a 2-mile contact, obscures the original magmatic relationships.

In three places on the Indian Creek side of the contact, detached from the Rosalie lobe, narrow short dikes of red dense aplite studded with milky quartz blebs are almost identical with known apophysal dikes within the main Pikes Peak batholith. These aplite stringers in the

Indian Creek granite, and the smooth cross-cutting edge of the Rosalie lobe against the Indian Creek and other formations, suggest that the Rosalie lobe of the Pikes Peak batholith may be younger than the Indian Creek plutons. However, the dikes in the Indian Creek granite which resemble Pikes Peak aplite may have come from residual magma long unconsolidated within the Pikes Peak intrusion which entered the fractures in the Indian Creek plutons and closely related gneiss and schist after the main bodies of Pikes Peak granite and the Indian Creek plutons were mostly consolidated. Some of the Pikes Peak granite may have been remelted during the period of stress and strain incident to the intrusion of the Indian Creek plutons. and this melt may have filled discrete fractures on the Indian Creek side of the boundaries.

Despite the somewhat contradictory record close to the contact of the two granites, the influence of the Rosalie lobe on the emplacement and distribution of the Indian Creek plutons definitely indicates that the Rosalie lobe is considerably older than the plutons.

INDIAN CREEK PLUTONS

GEOMORPHOLOGY

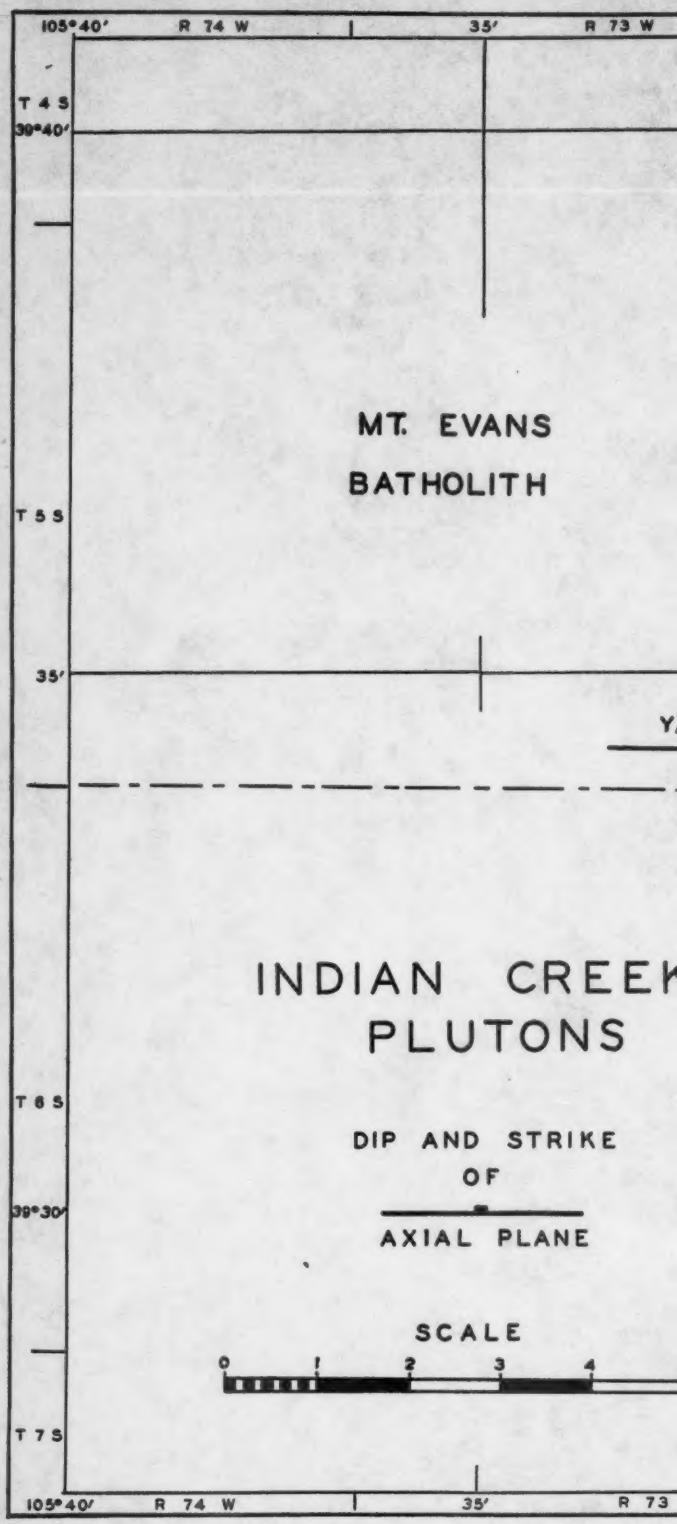
Where the four belts of plutons crop out concentrically along the northern and eastern margin of the Rosalie lobe of the Pikes Peak batholith, there are mountain masses comparable to the size of the granite bodies. Where most of the plutons have their long directions from northwest to southeast, the mountain contours are elongated in that direction.

Schist and gneiss of the Idaho Springs formation floor the flatter parts of most of the valleys between plutons and underlie the open mountain "parks." Main drainage lines are antecedent to the unroofing of the plutons from their coverings of schist and gneiss. Turkey, Cub, and Bear creeks, as well as other streams whose courses lie athwart the pronounced northwest—southeast grain of the country (Pls. 2, 3), have been adjusted along fault and sheared zones.

At the east, tributaries to main drainage are mostly consequent and follow the northwest—southeast bedding and foliation of the Idaho Springs beds. Insequent drainage is radial from the summits of plutons and locally follows the northeast—southwest foliation and joint systems of the coarse Indian Creek granite. Recent uplift of the region at the southwest has steepened the northeast slopes and renewed erosion on them.

At the west, the drainage is mostly radial from the massive peaks at the junction of the quartz monzonite of the Mt. Evans batholith, the Pikes Peak granite, and the Yankee Creek pluton of Indian Creek granite.

Where erosion has completely removed the schist and gneiss from the granite, its primary magmatic structures dominate the topography.


105°40'
R 74 W
35'
R 73 W
T 4 S
39°40'
MT. EVANS
BATHOLITH
T 5 S
35'
INDIAN CREEK
PLUTONS
T 6 S
DIP AND STRIKE
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T 7 S
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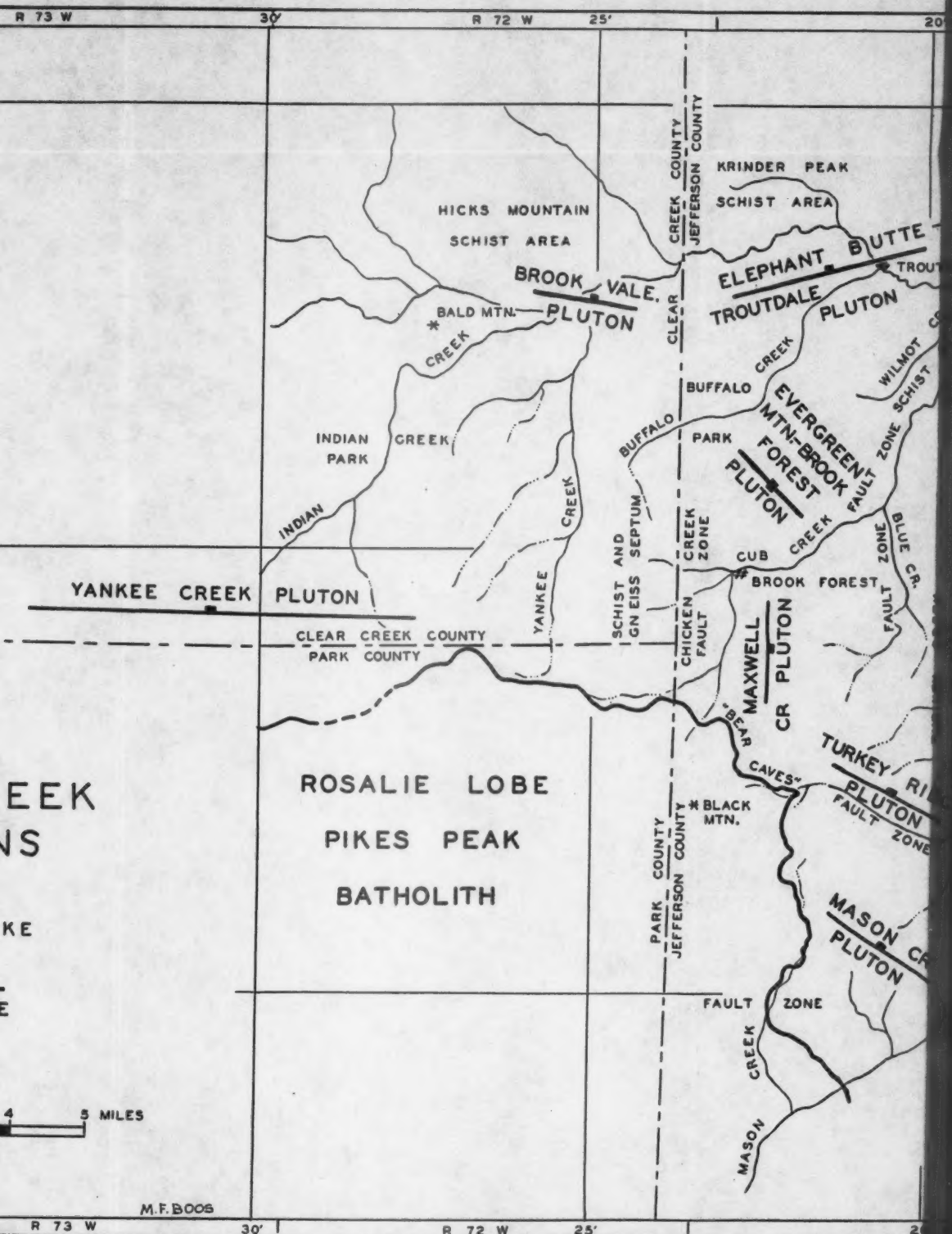
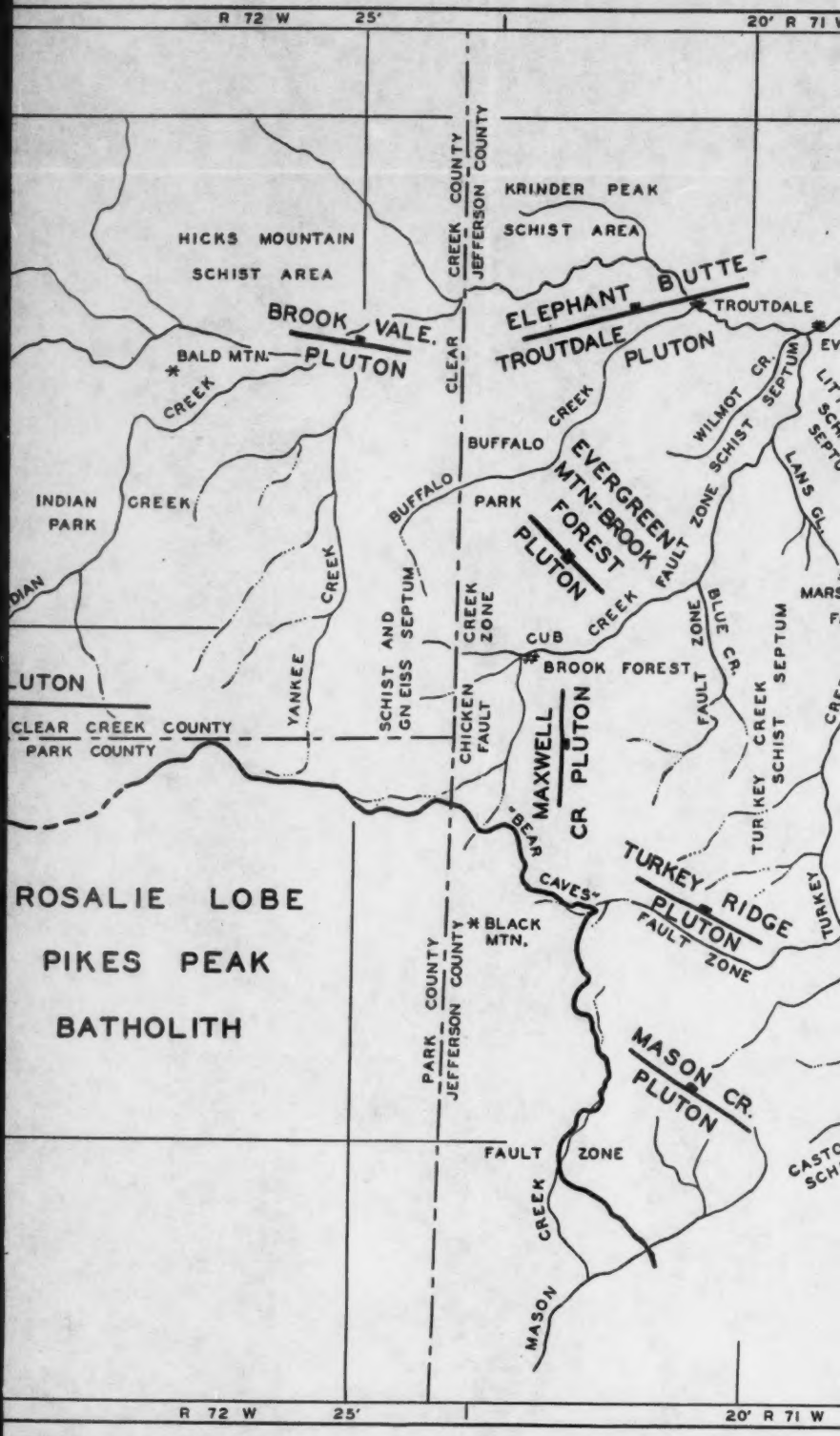
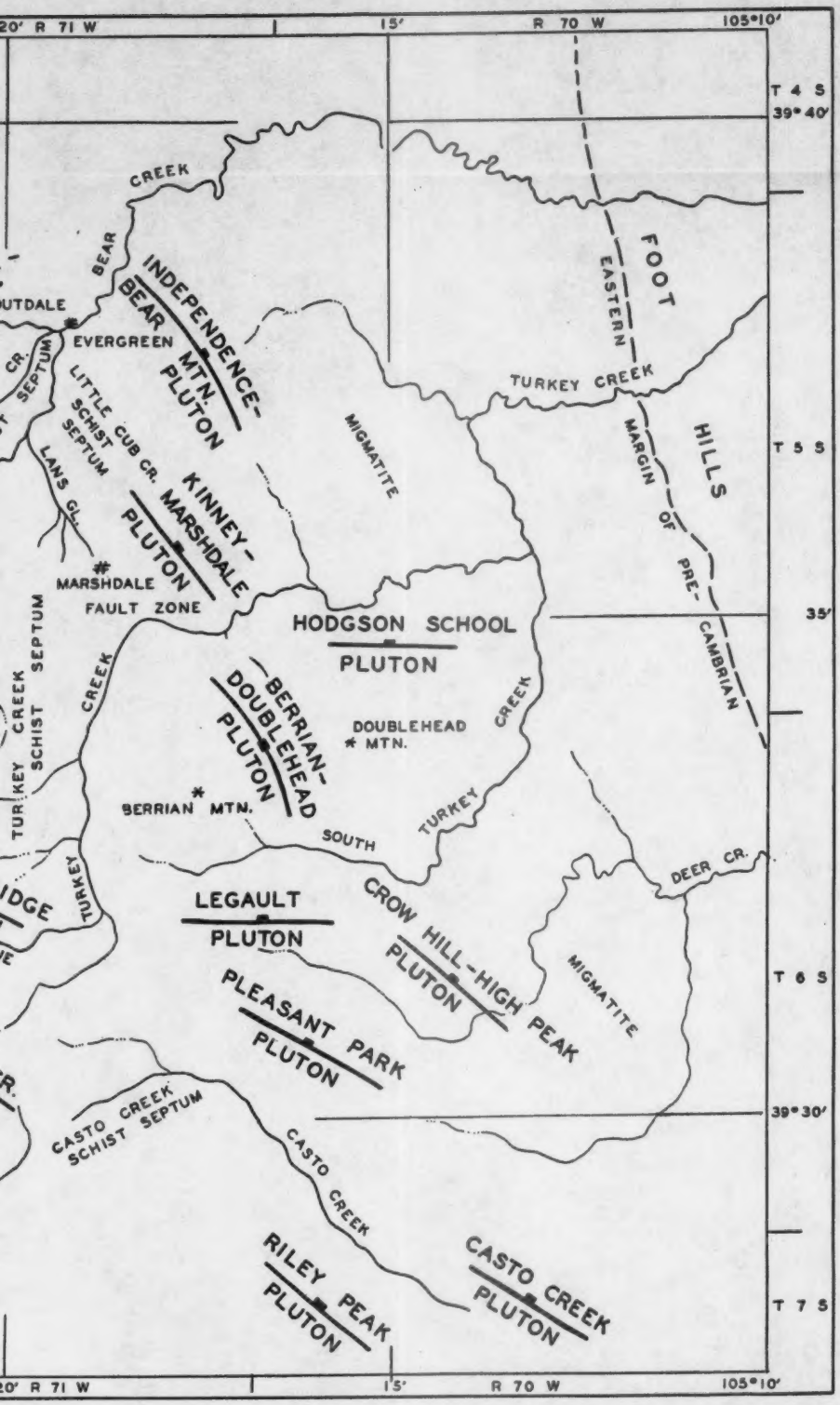
R 73



GEOGRAPHIC DISTRIBUTION OF THE INDIAN CREEK PLUTONS AND THEIR AXIAL



IC DISTRIBUTION OF THE INDIAN CREEK PLUTONS AND THEIR AXIAL TRENDS


20' R 71 W
15'
R 70 W
105°10'
T 4 S
39°40'
T 5 S
35'
T 6 S
39°30'
T 7 S
BEAR CREEK
INDEPENDENCE-BEAR MTN. PLUTON
EVERGREEN
LITTLE CUB CR. SCHIST SEPTUM
LANS GL.
KINNEY-MARSHDALE PLUTON
MARSHDALE FAULT ZONE
MIGMATITE
FOOT HILLS EASTERN MARGIN OF PRE-CAMBRIAN
TURKEY CREEK
HODGSON SCHOOL PLUTON
BERRIAN-DOUBLEHEAD PLUTON
DOUBLEHEAD MTN.
BERRIAN MTN.
TURKEY CREEK SCHIST SEPTUM
SOUTH TURKEY CREEK
DEER CR.
LEGAULT PLUTON
CROW HILL-HIGH PEAK PLUTON
PLEASANT PARK PLUTON
CASTO CREEK SCHIST SEPTUM
CASTO CREEK
RILEY PEAK PLUTON
CASTO CREEK PLUTON

TRENDS

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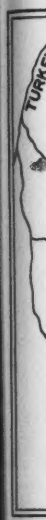

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Cupolas, spines, and dikes (intraplutonic and peripheral) are topographic features. Knobby granite bosses mark centers from which granite magma welled upward, and, at the northwest, the plutons terminate in long subparallel ridges of granite which fade out into gneiss and schist (Figs. 2,

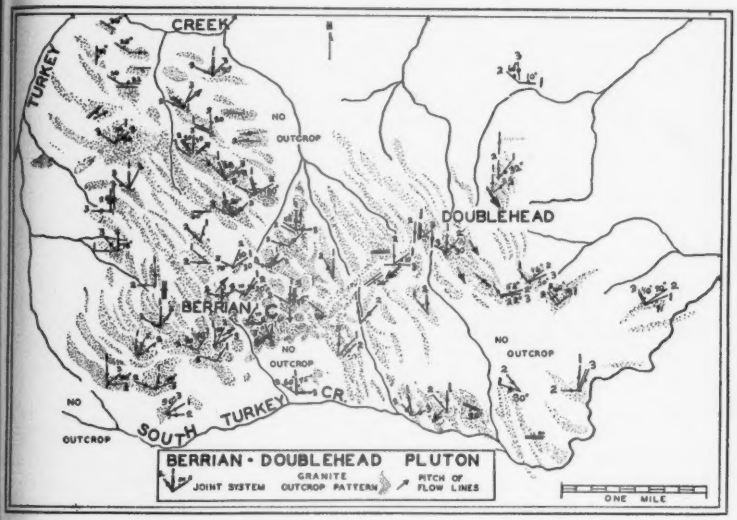


FIGURE 2.—*Structural and topographic features of the Berrian-Doublehead pluton*

8, 10; Pl. 4). The southern termini of many plutons show fault line cliffs where the granite was faulted up from the adjacent schist and gneiss.

NUMBER AND ARRANGEMENT

Sixteen plutons of Indian Creek granite have been identified; each has developed from either a single center of magmatic action or from a group of closely related centers. The arrangement of the plutons in the pre-Cambrian terrane of the region is in no way haphazard. They are disposed along definite trends, and each is, mostly, an integral structure. All lie peripheral to the Rosalie lobe of the Pikes Peak batholith and crop out in four semicircular arcs (Fig. 1; Pl. 2). Most plutons are elliptical in outcrop pattern and have podlike or lenticular bodies. Both long and short axial directions have been discerned (Fig. 3).

Along the margin of the Rosalie lobe of the Pikes Peak batholith there is an almost continuous outcrop of Indian Creek granite. Structurally, it consists of the Yankee Creek, Maxwell Creek, Turkey Ridge, and Mason

Creek plutons. This inner semicircular belt of plutons is almost free from cover or roof pendants. The contact of each pluton with its neighboring pluton, or other wall rock, is definitely traceable although very little schist and gneiss separate the plutons.

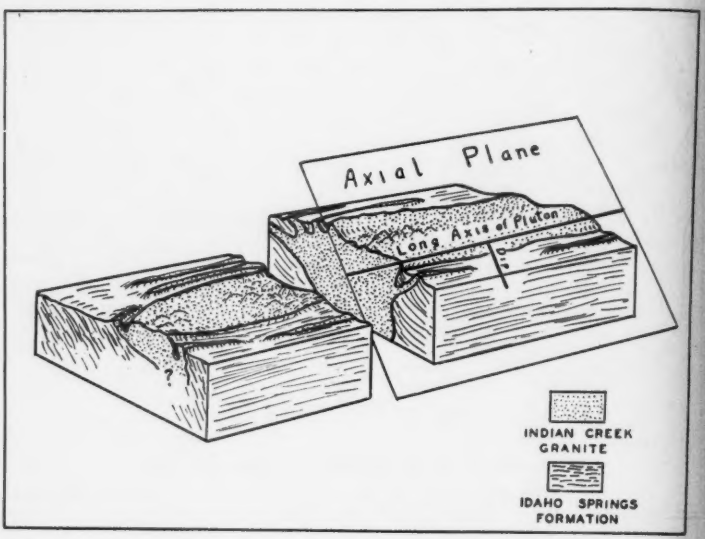


FIGURE 3.—*Diagrammatic representation of a pluton showing the axial plane and the axial trend*

The Brookvale, Brook Forest-Evergreen Mountain, Riley Peak, and Casto Creek plutons form a concentric granite belt parallel to the inner arc of plutons but separated from it by well-defined septa of migmatite and schist. Long narrow bodies of schist and gneiss separate the plutons of the second set, and patches of migmatite and schist film the granite summits.

The third belt of plutons—Kinney-Marshdale, Berrian-Doublehead, Legault and Pleasant Park—parallels the first two. Along Blue and Turkey creeks north of Conifer and in the Wilmot Creek-Lans Gulch areas irregular septa of schist and gneiss separate the third row of plutons from the second.[2]

The fourth and outermost set of plutons includes the Independence-Bear Mountain and High Peak-Crow Hill bodies. Contacts of plutons

[2] This alignment and association of plutons differs from those given in an earlier paper (Boos, 1936). Recent detailed field studies of structure and topographic trends warrant the new arrangement.

FAULT AND DIKE PATTERN OF THE INDIAN CREEK PLUTON AREA

LEGEND

MARGIN OF ROSALIE LOBE OF PIKES PEAK BATHOLITH

FAULTS

PEGMATITE DIKES

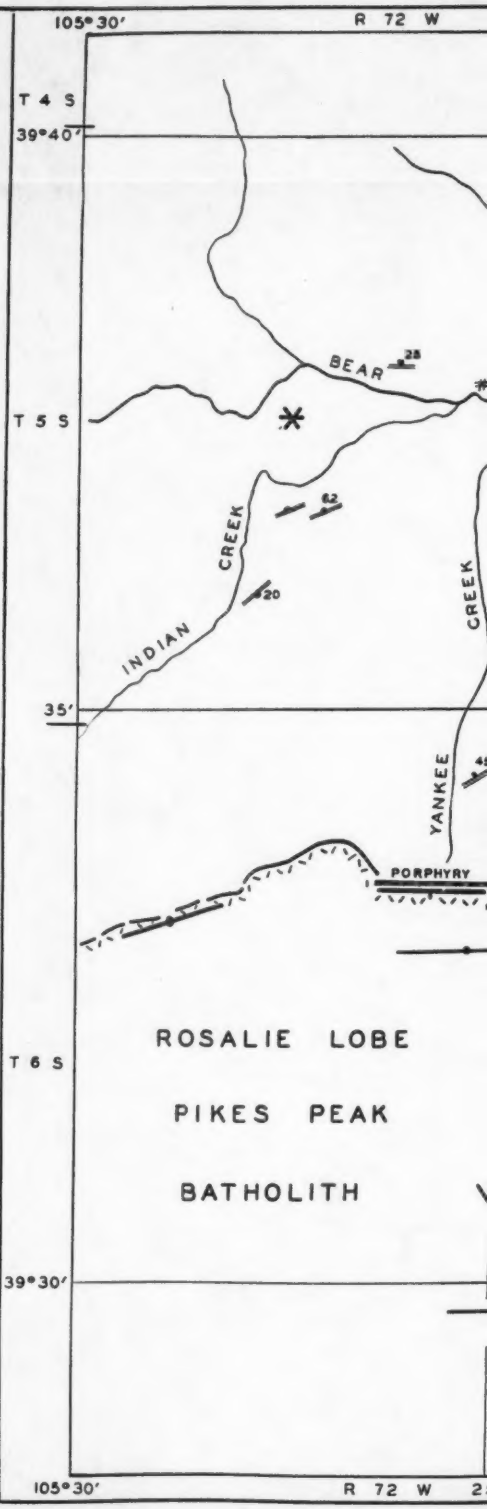
QUARTZ VEINS

CENTERS OF ACTION IN PLUTONS

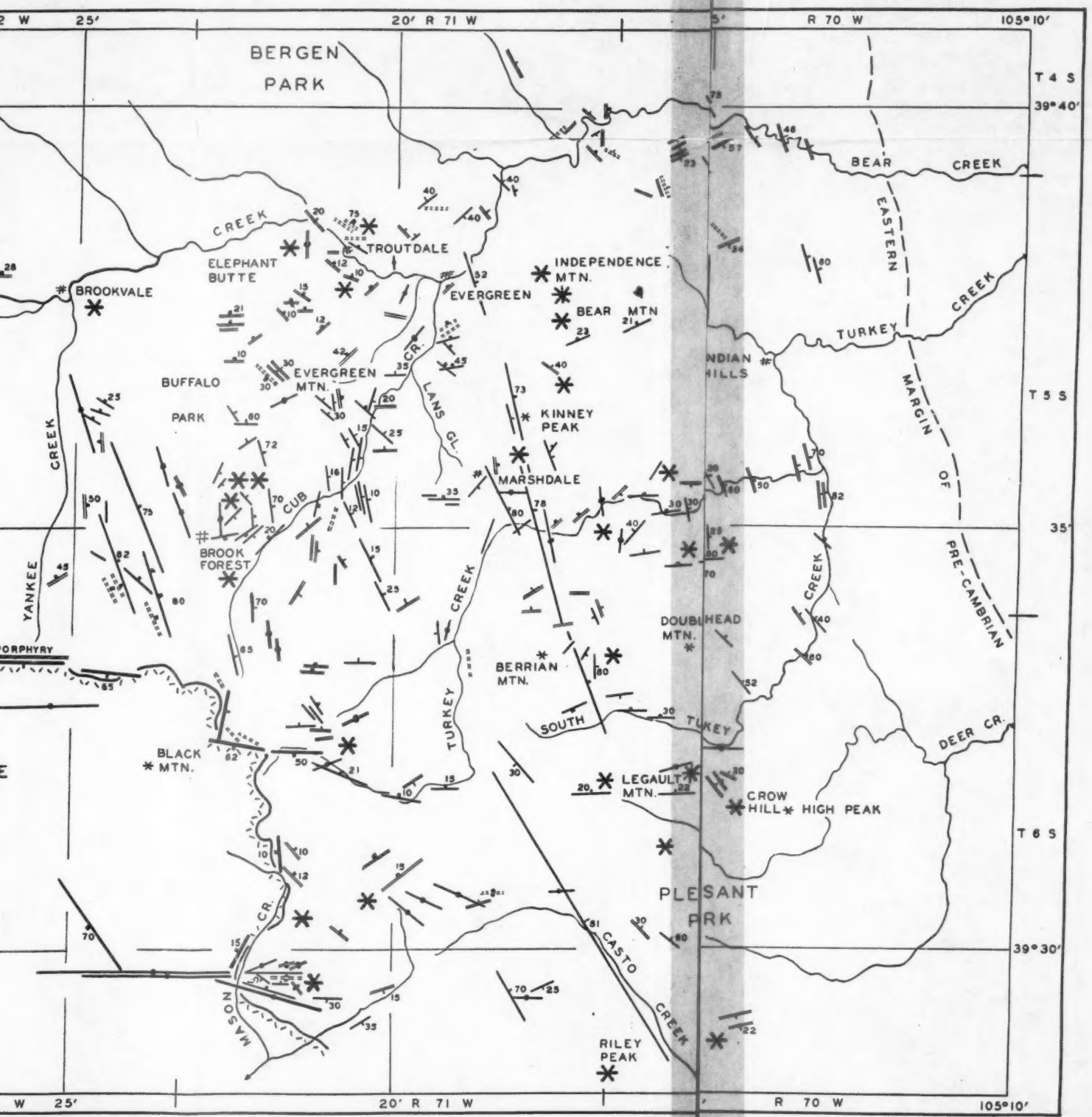
SCALE

0 1/2 1 2 3 4 5 MILES

GEOLOGY BY M. F. BOOS AND ASSISTANTS



FAULT AND DIKE STRUCTURES



CTURES IN THE INDIAN CREEK PLUTONS AREA

against the Idaho Springs formation are everywhere visible at the northeast and north.

A few plutons depart from the general subparallel arrangement about the margin of the Rosalie lobe. The Elephant Butte-Troutdale pluton lies athwart the general northwest—southeast trend of the others, its long axis extending from northeast to southwest. The long axis of the small pluton south of Hodgson School lies east—west, directly across the general north—south trends of schist and gneiss foliation in the outer row of plutons.

North and northeast of the outer row of plutons few bodies of granite other than dikes and sills crop out in the Idaho Springs beds. There are no structures reflected in the Idaho Springs beds north and east of the plutons described which would suggest the presence of large granite masses relatively near the surface.

TYPES

General statement.—Three types of plutons are discriminated in the region; the basis for determination of the type is the number of centers of action present within the pluton unit. The phrase "center of action" is used to indicate a focal area where primary structures reveal a channel of upwelling and deploying magmatic currents. Structural details of such centers are discussed under the Berrian-Doublehead pluton.

Uninuclear.—The primary structures of individual plutons of this type are closely related to one center of magmatic activity and deployment. Growth of the pluton proceeded from a single source of magma, and the peripheral beds were invaded lit-par-lit by sills.

The Maxwell Creek, Turkey Creek, Riley Peak, and Casto Creek plutons are uninuclear. The Berrian-Doublehead pluton is typically uninuclear and has been mapped in detail. All parts are well exposed and accessible from road and trail.

Topographically, the Berrian-Doublehead pluton shows, at the north, two parallel high ridges of north—south to N. 40° W. trend connected at the south above 8000 feet by an east—west ridge marked by many bare knobs of granite. The valleys of north and south Turkey Creeks, surrounding the pluton, are mostly underlain by schist and gneiss. By a number of piracies, not detailed here, the streams were brought to follow the foliation of schist, gneiss, and sheared zones, some of which were of magmatic origin. The entrenchment of the drainage lines has resulted in the removal of the enclosing schist and gneiss, and the granite stands out in bold relief (Fig. 2; Pl. 5).

The pluton is an asymmetrical, lenticular body, with its long axis northwest—southeast and a nearly vertical axial plane. The one well-

defined center of magmatic activity is exposed on the low ridge which links Berrian and Doublehead Mountains (Pl. 6). This core of the granite body is distinguished, topographically, by many domical erosion

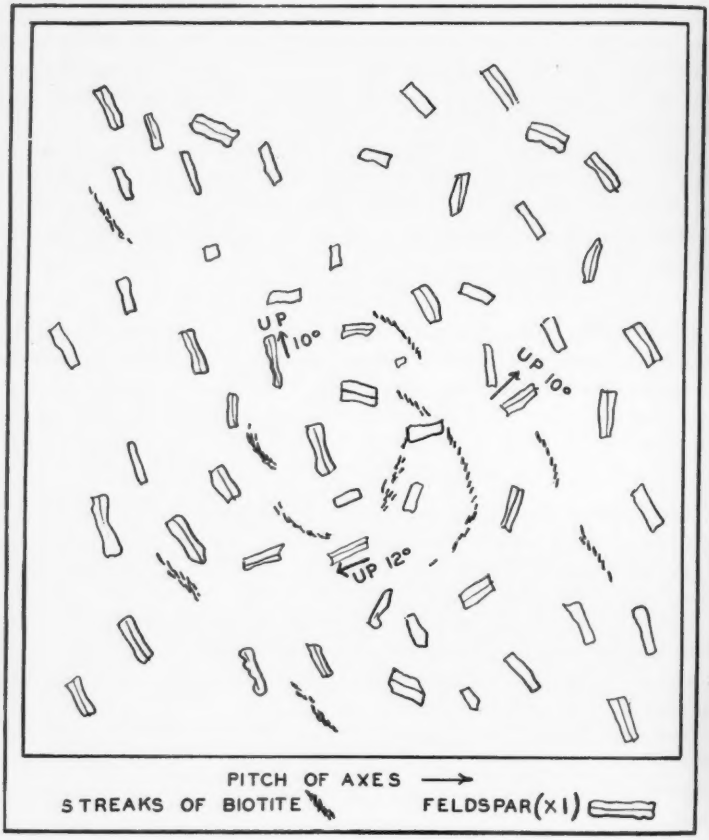


FIGURE 4.—*Swirl pattern of long tabular feldspar crystals*

Area shown is about 10 square feet; the feldspars are enlarged although spacing is kept relatively correct.

forms; knobs, thumbs, and bosses of granite surmount lumpy ridges. The core is composed of coarse trachytoid granite and yields to weathering more readily than the dike and sill complexes of Berrian Mountain on the west and Doublehead Mountain on the east flanking the mass of the pluton (Pl. 5, fig. 2).

FIGURE 1. SOUTHEAST END OF BERRIAN MOUNTAIN AND PART OF CENTER OF ACTION ON EAST FLANK OF MOUNTAIN
Flat valley of Turkey Creek in foreground; view is N. 25° W.

FIGURE 2. SOUTHERN END OF SMALL GRANITE BODY WEST OF HODGSON SCHOOL
View is N. 15° W.

TOPOGRAPHIC DETAILS OF SOUTHERN ENDS OF TWO PLUTONS

FIGURE 1. VIEW EAST FROM BLACK MOUNTAIN
(A) Summit of Berrian Mountain; (B) long smooth northwest slope; (C) Turkey Ridge pluton; (D) Riley Peak pluton.

FIGURE 2. BERRIAN MOUNTAIN (B) FROM WEST SLOPE OF DOUBLEHEAD MOUNTAIN
Low ridge (A) marks center of action of pluton structure.

FIGURE 3. VIEW N. 20° E. FROM BERRIAN MOUNTAIN
(A) low ridge of granite marking center of action; (B) Doublehead Mountain.

At the core, or center of action, the long axes of the tabular orthoclase feldspars are arranged in swirling patterns, and their orientation, viewed on a nearly horizontal surface, is shown in Figure 4. If the parallel tabular feldspars and the faint streaks of mica in the coarse granite

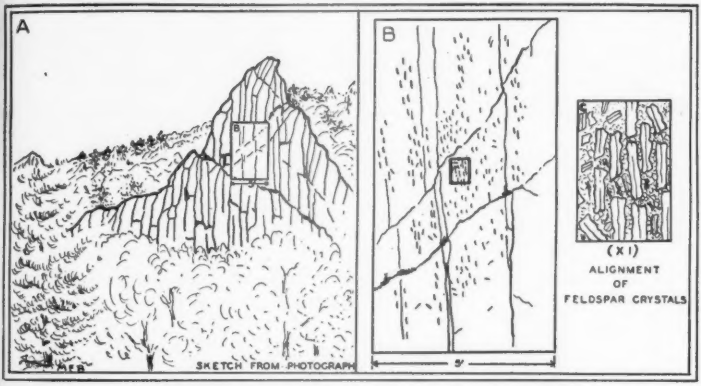


FIGURE 5.—*Orientation of tabular feldspar crystals on weathered surface of spiny granite knob on the core of the Berrian-Doublehead pluton*

A. Appearance of field outcrop; note the well-defined joint pattern.
B. Detail of A with pronounced vertical alignment of the long tabular feldspars.
C. Small section of structures shown in B with vertical position of the twinned long tabular feldspar (microcline) crystals sketched.

indicate magmatic flow, their funnel-like orientation and deployed positions suggest that the currents of magma spread from one central orifice compounded of many swirling currents. The general upwelling of fluids and their spread from the single core built the pluton. The core itself was not a single funneling conduit, but the currents within it were confluent and blended from the center of action. Flow lines and flow layers within the core indicate the general convectional movements of the magma, but away from the core area the platy structures strike N. 40°–60° W. in common with the structural trends of the surrounding schist and gneiss.

In addition to the funneling structures there are zones of vertical foliation in which the long axes of the tabular feldspars are vertical. These structures lie in the midst of granite which shows more or less haphazardly arranged tabular elements (feldspar and mica). The knobs and pinnacles on the core exhibit the vertical foliation planes with vertically disposed feldspars and strike mostly northwest—southeast. The funneling structures rarely crop out in knobs and spines. They are mostly exposed in low places, where erosion has planed the granite to a horizontal surface. The granite knobs and pinnacles on the core are

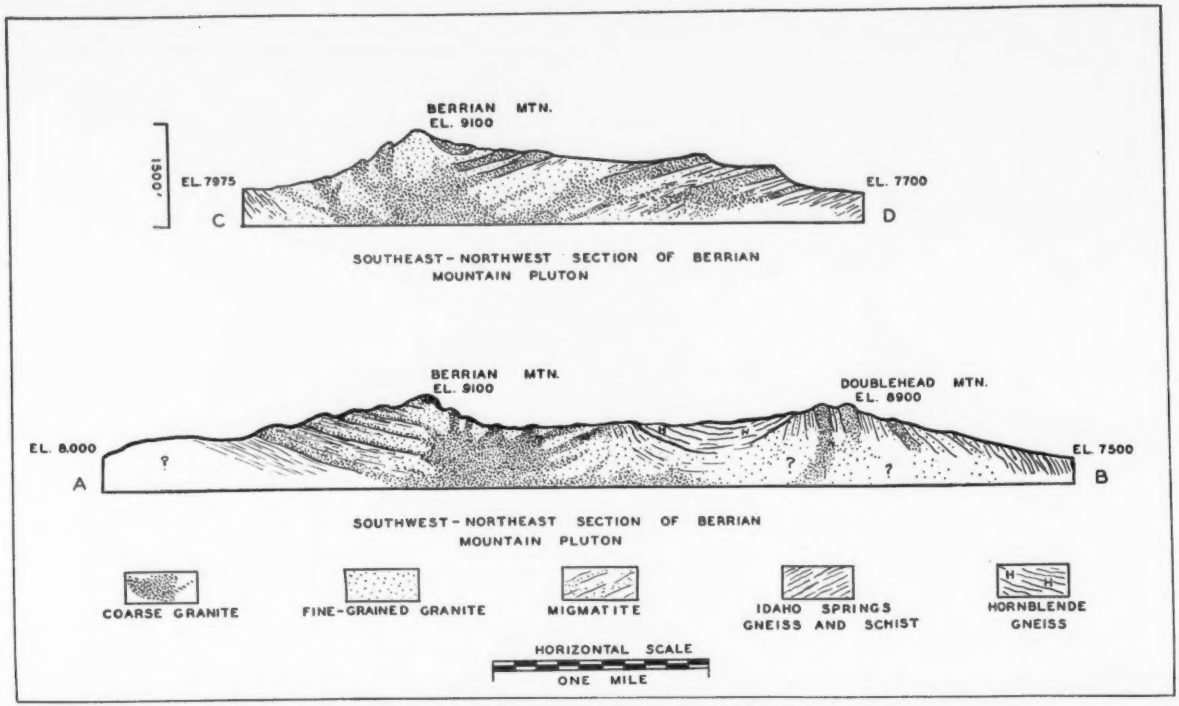


FIGURE 6.—*Sections across the Berrian-Doublehead pluton*

the topographic expression of magma movements which carried the slush of feldspar crystals into a nearly vertical position. The vertically disposed feldspar crystals weather out conspicuously on the flanks of the pinnacles (Fig. 5). Where they are not part of the funnels and swirls and vertically foliated masses, the long axes of the feldspars pitch down to the north and south away from the center of action; this pitch near the periphery of the core is probably a drag phenomenon due to friction exerted by the inert walls of the pluton against the rising and spreading fluids. The invasion of the magma as a whole was more permissive than aggressive, and the currents rose generally parallel to the schistosity of the enclosing country rock.

South of the center of action, the pluton is sheared off by east—west to northwest—southeast faults which have dropped the Idaho Springs beds below the level of the granite core.

Although the granite at the core has been mostly unroofed, thin patches of infolded schist and gneiss are preserved in some of the low places, and their downfolded structures indicate that the roof was accommodated to the irregular surface of the core, being arched over the protrusions and depressed between the cupolas. In the few places where a section is exposed from roof down into coarse granite, there is a zone of migmatite between the coarse granite of the core and the schist of the roof (Pl. 9, fig. 2). The streakiness of migmatite trends mostly east—west and dips to the north at a low angle. Contact between the migmatite and the schist and gneiss of the roof strikes northwest—southeast and dips northeast at a low angle.

Bordering the core on the east and west, lenses and masses of coarse trachytoid granite strike northeast—southwest and dip northwest at a low angle. They transect the fine-grained migmatite which lies below the roof. It appears that the migmatite represents the earliest phase of pluton intrusion, developed when granitic fluids were just beginning to penetrate the area, while the trachytoid granite core represents the later phase formed when the magma continued to move into an area thoroughly saturated with and heated by the earlier fluids.

Flanking the core on the east and west are the large granite and schist masses of Berrian and Doublehead mountains. Berrian Mountain consists of many thick lenses of coarse granite in schist and gneiss. Lower down on the western slope of Berrian, the schist and gneiss beds strike north—south and dip to the east at a low angle. Higher on the slopes the lenticular bodies of granite and the intervening schist beds retain their north—south strike, but the eastward dip steepens toward the center of the pluton. It is believed that the sills were derived from and are in connection with the granite core (Fig. 6). The summit of the mountain

FIGURE 6.—*Sections across the Berrian-Doublehead pluton*

FIGURE 7.—*Tectonic map of the Berrian-Doublehead pluton*

is filmed with schist and migmatite which strike east—west and dip north at a low angle. All thick granite sills on the western side of the mountain show gradation with elevation from fairly coarse granite below to a fine-textured rock containing migmatite and finally schist and gneiss in the roof. This sequence is repeated nearly to the top of the mountain.

The eastern side of the core is concealed by foliated hornblende gneiss which trends N. 60° E. (Fig. 7). East of the hornblende gneiss, on Doublehead Mountain, there is a large body of platy migmatite whose foliation trends mostly northwest—southeast and dips southwest. The double summit of the mountain consists of lensing masses of granite which cut through the migmatite and schist of the flanks and lie mostly N. 10° W. and dip steeply southwest. The eastern side of the pluton is sheathed in migmatite and schist, where locally knobs and cupolas of Indian Creek granite pierce the cover of metamorphic rocks.

Northward from the core area, north—south-trending granite ridges appear to be dikelike offshoots from the center of action. Near the core, the dikes are coarse granite, but the granite is finer grained distally and grades into very fine-grained rock containing abundant migmatite. The dikes show migmatite full of schist inclusions at the northwest termini.

A valley 500 feet deep on the north flank of Berrian Mountain shows a section of the pluton below the roof area. At the base of the steep cliffs there is coarse trachytoid granite with northwest-dipping flow layers. The coarse granite bodies are mostly free from inclusions and lack pegmatite and aplite dikes. The granite appears to be in lensing bodies which strike northeast—southwest and dip northwest. Their lower limits are not visible, but on their summits the lenses penetrate the migmatite roof. The migmatite and fine-textured granite which these coarse bodies cut is streaked with northwest—southeast-trending inclusions of biotite schist and shows foliation and platy parting which strikes northwest—southeast and dips steeply northeast.

The coarse granite nearly everywhere underlies the migmatite and schist or penetrates them from below. The primary structures of the migmatite and fine-grained granite of the upper parts of the pluton trend northwest—southeast and dip northeast, but the flow layers of the coarse, trachytoid, granite trend northeast—southwest and dip northwest.

The Berrian-Doublehead pluton is an integral structure passively introduced lit-par-lit within the surrounding schist and gneiss from a core or center of action. The pluton appears unrelated, except by position in the general pattern of plutons, to neighboring bodies of similar granitic makeup and structural pattern. The pluton may, however, be a summit portion of a much larger subjacent body, perhaps of batholithic propor-

tions. The conclusions, drawn mostly from field evidence, are that the Berrian-Doublehead pluton is an individual unit composed of closely interpenetrated and coalesced magma currents originating in a core or center of action. Marked flow structures reveal the direction of these

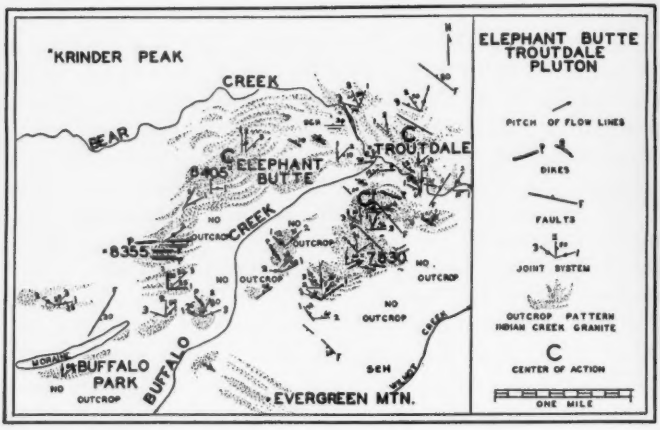


FIGURE 8.—*Structural and topographic features of the Elephant Butte—Troutdale pluton*

currents. The great difference in the direction of flow in the upper and lower levels of the pluton mass is ascribed to a change in the regional stresses during the successive stages of intrusion. It may be that the size and shape of the space available for granite intrusion varied in response to an interplay of magmatic and regional stresses produced concurrently with emplacement; the flow of magma being directed into areas of least resistance which were slowly shifted as stresses fluctuated in intensity and shifted in orientation as the intrusion progressed.

Complex.—Some plutons resulted from magma currents originating in several centers of action whose streams became so interpenetrated and blended that they made one plutonic mass. Each complex pluton is a single unit of intrusion within the surrounding Idaho Springs gneiss and schist and is set off from adjacent granite plutons by well-defined septa of schist and by fault and shear zones. The Elephant Butte-Troutdale, Brook Forest-Evergreen Mountain, Brookvale, Legault, Crow Hill-High Peak, Pleasant Park, and Hodgson School granite bodies are complex plutons.

The Elephant Butte-Troutdale pluton includes a long massive granite lens with a bulbous eastern head (Pl. 7) which narrows to the southwest,

FIGURE 1. GIANT SPINE OR THUMB OF GRANITE ON EAST FACE OF BERRIAN MOUNTAIN
View is to north along road into the Tony Granzella Ranch. This is the south face of outcrop sketched in Figure 5, which shows the north face of thumb.

FIGURE 2. LUMPY KNOBS ON THE CORE OF THE BERRIAN-DOUBLEHEAD PLUTON
View is N. 30° W. from the east side of the center of action.

TOPOGRAPHIC DETAILS OF CORE AREA OF BERRIAN-DOUBLEHEAD PLUTON

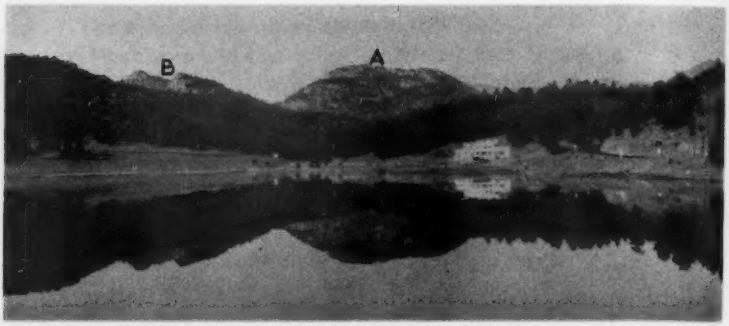


FIGURE 1. ELEPHANT BUTTE (A) AS SEEN FROM EAST END OF EVERGREEN LAKE
Note rounded blunt outcrops of east face of the Butte; lumpy domical knobs (B) of center of action south of the Butte.

FIGURE 2. EASTERN END OF ELEPHANT BUTTE DETAIL OF (A) IN FIGURE 1
View is to northwest and reveals the characteristic jointed outcrop of Indian Creek granite on its bulbous eastern head.

FIGURE 3. ELEPHANT BUTTE PLUTON FROM THE NORTH
(A) Same place as (A) of Figure 1; (B) is (B) of the same figure; (C) Riley pluton; and (D) Berrian-Doublehead pluton.

OUTCROPS OF INDIAN CREEK GRANITE ON THE ELEPHANT BUTTE-TROUTDALE PLUTON

grading into migmatite and schist and ending in comb ridges of schist which dip to the northwest steeply and a cluster of lumpy domical granite knobs lying south of the head area and rising steeply above the schist under the flat floor of Wilmot Creek valley. Main drainage lines do not

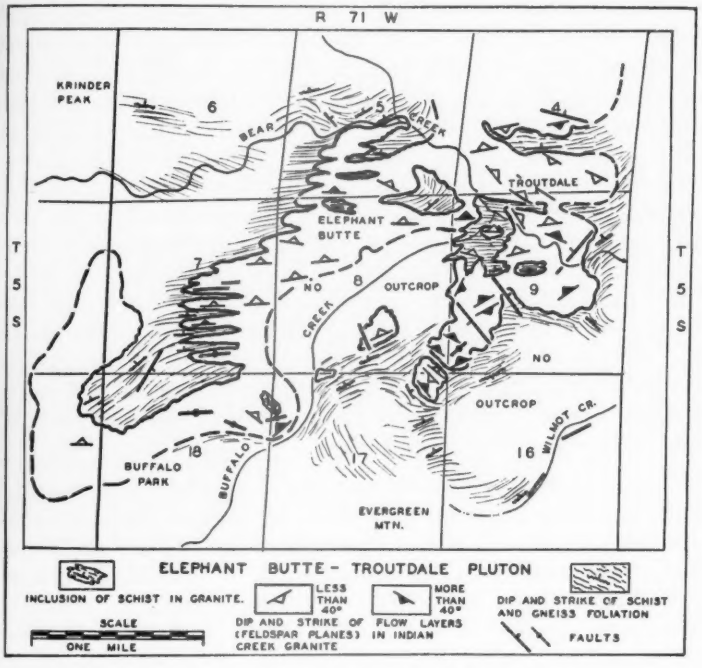


Figure 9.—Tectonic map of the Elephant Butte-Troutdale pluton area

generally cross plutons, but Bear Creek has trenched the area between Elephant Butte and Troutdale, and Buffalo Creek meanders across the southern complex of core structures (Figs. 8, 9).

The long structural axis of the pluton as a whole trends N. 60° E., and the axial plane dips steeply northwest. The granite appears to have spread from three centers of action: (1) At the east end of Elephant Butte; (2) north of Troutdale, and (3) among the domical hills between Troutdale and Wilmot Creek (Fig. 8). North of the pluton the foliation of the Idaho Springs beds dips northward. On the south side of the granite body, the gneiss and schist beds dip southward and are a part of the syncline which underlies Wilmot Creek valley.

The pluton is an asymmetrical and irregular anticlinal body. Drag folds show in the schist and gneiss of its north flank. The pluton is centered about core areas in which flow structures follow planes striking N. 60° E., with the linear elements plunging to the southwest. The steeper flanks of the pluton are at the northwest. Except where glaciation has levelled the outcrop under Buffalo Park, the topographic surface closely follows the tops of the domical centers of intrusion. The big granite laccolithlike blister exposed in vertical section on the cliffs north of Troutdale and the lumpy granite body at the eastern end of the Butte mark cores from which magma spread mostly lit-par-lit into gneiss and schist. The center of action south of the Butte exhibits funneling and swirling flow pattern of tabular feldspar like those within the core of the uninuclear Berrian-Doublehead pluton. Dikes radiate from the cores, and sills spread from the same parts of the plutons, except at the south and southwest where the pluton lacks apophysal structures.

Schist and gneiss underlie, at least locally, the granite of the Butte and the Troutdale centers of the pluton. Thus it appears that part of the pluton is laccolithic; the granite deployed through a floor from two or three centers of action, which pierced the basal schist and made a "pine tree" pattern in which the granite lenses overlap and are imbricated within the body of the pluton.

Compound.—Some plutons have widely spaced nuclei or centers of action, and the primary structures of the whole granite body are related to the several cores. The Bear Mountain-Independence Mountain and the Kinney-Marshdale plutons are compound. In the compound plutons, although the magma spread from the various centers of action individually, the granite streams are mostly confluent marginally and form a compound body set off from its neighboring plutons by wide septae of schist and gneiss whose foliation and bedding planes strike northwest—southeast and dip northeast. Small bodies of schist and gneiss separate the centers of action within the compound pluton, and locally the bedded schists and gneisses curve over the upper limits of granite masses.

The Kinney-Marshdale pluton has a very irregular topographic pattern (Fig. 10) closely related to the compound structures developed from several widely spaced centers of magmatic action. Generally, cores of the pluton mass make topographic highs, and septae of schist and gneiss underlie the valleys.

There are three well-defined centers of action: (1) On the eastern flank of Marshdale Peak, (2) in the northeastern lobe of the pluton, and (3) among the domical hills at the extreme southwestern part of the pluton

(Fig. 10). The granite of all three shows the swirl and funnel structures which characterize the cores of other plutons in the Indian Creek area. The cores are festooned with dikes and sills which extend radially from the central areas and reinforce the adjacent schist and gneiss and

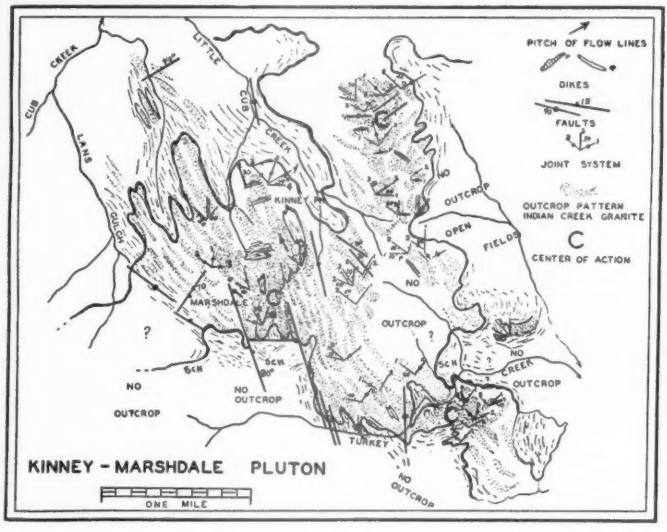


FIGURE 10.—*Structural and topographic features of the Kinney-Marshdale pluton*

migmatite wall rock. Locally, the apophysal bodies have been faulted out so that the cores appear isolated (Fig. 11).

On the south side of the pluton, the granite masses are bulbous and blunt and thin out to the north and northwest lit-par-lit into beds of schist and gneiss. The small center of action at the southeast is shut off, topographically, from the main body of the pluton by the valley of Turkey Creek. The domical hills and knobby ridges trend mostly northwest—southeast paralleling the primary structures in the granite. Granite and migmatite masses are cut off, at the south, both structurally and topographically, by the fault zone northwest of Doublehead Mountain.

Southwest of Marshdale there may be a fourth center of action which belongs to the Kinney-Marshdale pluton. Scattered granite outcrops show a partly exposed core, but this center may belong to a small pluton which is barely reflected in the schist between Blue and Turkey creeks (Pl. 3).

The northeast lobe of the Kinney-Marshdale pluton is plainly laccolithic. Along several small valleys a floor of schist and gneiss is shown below the granite which spreads and widens upward making a center

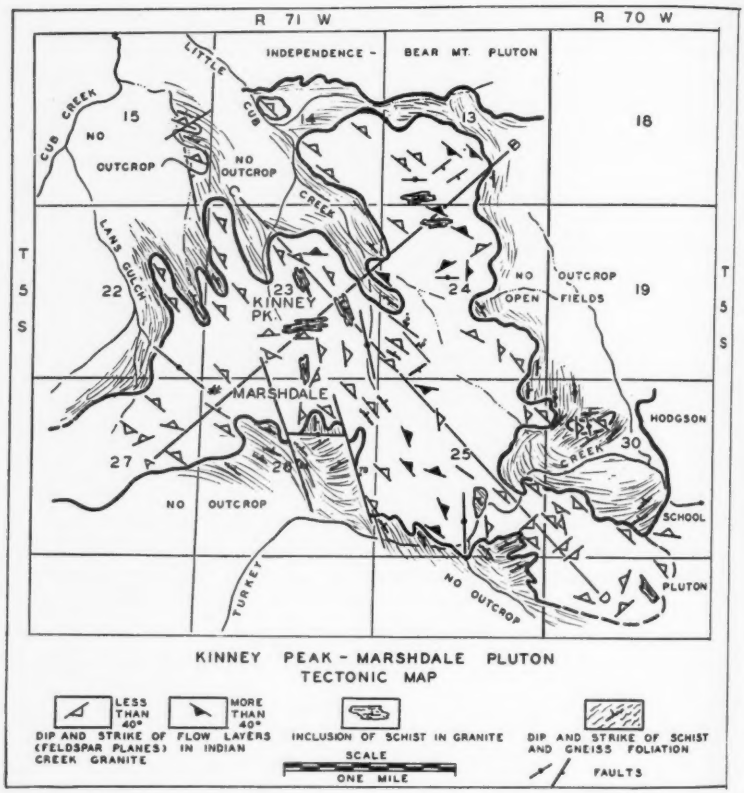


FIGURE 11.—*Tectonic map of the Kinney-Marshdale pluton area*

of deployment below the apical part of the structure. Locally, schist and gneiss beds curve over the roof areas (Fig. 12).

On the summit of Marshdale Mountain a film of schist and gneiss shows foliation (probably original bedding of the sedimentary rock) dipping north at a low angle. Mostly, there is no migmatite selvage between the coarse granite of the core areas and the lower parts of the pluton, and the roof structures of schist and gneiss. Locally, fine-grained pepper-

and-salt granite underlies the schist and gneiss, but in most exposures the contact of the coarse granite against the metasedimentary rock of the roof is well defined. The bedding (foliation) of most of the roof of metamorphic rock strikes northwest—southeast and dips steeply northeast.

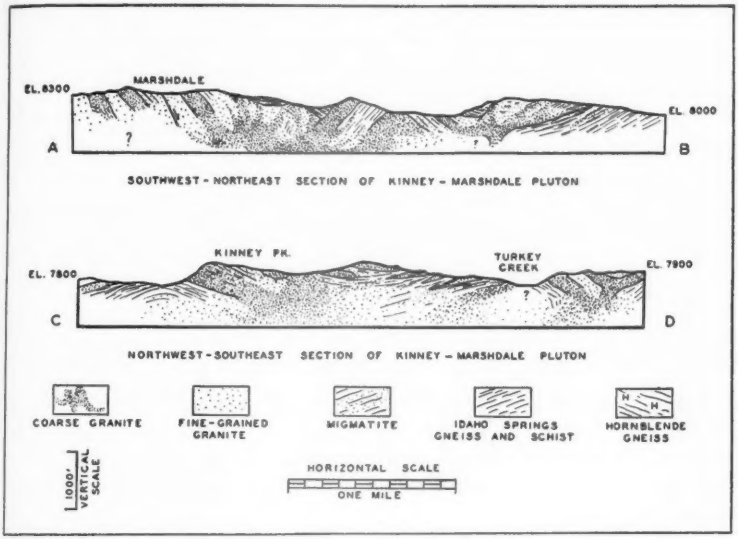


FIGURE 12.—*Sections across the Kinney-Marshdale pluton*

The coarse trachytoid granite of the Marshdale core has tabular feldspars (microcline) with the long axes striking N. 40°-60° E. The platy flow planes strike N. 40°-50° E. and dip northwest at high angles. To the east of this core the roof contains many small bodies of granite lit-par-lit with schist and gneiss. These podlike small granite bodies are blunt and thick at the south end and lense down into sill-like stringers within the schist at the northwest (Fig. 13). Below, close to the coarse granite of the core, the granite bodies are parallel and almost confluent laterally, but individual lenses are generally enclosed in thin wrappings of mica gneiss. In the upper parts of the roof, the "tadpoles" of granite are smaller, shorter, and thinner and lie isolated within the beds of schist and gneiss which accommodate them.

The granite bodies associated with the individual centers of action within the Kinney-Marshdale pluton might be considered small separate plutons. They are, however, grouped into one massive irregular

pluton because: (1) The pluton as a whole is set off from its neighboring granite plutons by fault and shear zones of considerable extent and magnitude. (2) The faults, which block in the compound pluton, are localized within the schist and gneiss bordering the massive bodies of granite, and many penetrate the granite body. (3) The roof structures are confluent—no marked septa of schist and gneiss divide the centers of action. (4) Well-defined septa of schist enclose the pluton and separate it from other adjacent plutons. (5) The granite masses of the cores have subparallel alignment and are similarly disposed.

For the reasons noted and as shown by the field record of internal and wall rock features, the various centers of action already described with their satellitic and apophysal structures belong subjacently to a single plutonic body—the Kinney-Marshdale pluton—a massive independent unit in the structural pattern of the Indian Creek area.

FEATURES COMMON TO THE PLUTONS

General statement.—The 16 plutons have many features in common, although each unit has some characteristic peculiar to itself. For clarity in comparison, these features have been tabulated (Table 2).

Distribution.—The four rows of plutons are concentrically and somewhat irregularly placed about the margin of the Rosalie lobe of the Pikes Peak batholith (Pl. 2), with an average of four plutons to a row and long narrow northwest–southeast-trending foliated schist bodies between the rows.

Size.—The plutons are relatively small granite bodies. The largest is about 12 square miles in area (Yankee Creek), the Berrian-Doublehead pluton is about 2 by 3 miles across, and the smallest pluton is half a mile wide by a mile long. The thickness of the granite is unknown; at least 2000 feet is exposed locally.

Shape.—The plutons have, generally, a lenticular to podlike form. They are not true laccoliths, although parts of many appear laccolithic.

Complex or compound plutons have a somewhat lobate surficial pattern. Generally, the bulk of the granite is asymmetrical with one massive thick end; the other parts pinch down into the surrounding and enfolding schist and gneiss. Generally, the blunt massive parts are at the south or southwest, but some plutons, such as Bear-Independence and Turkey Ridge, have the massive granite areas exposed at the northwest. The blunt thick part extends few apophyses into the surrounding schist and gneiss, and the outcrops have characteristic clifted profiles where the granite is faulted against the enclosing metamorphic rocks (Pl. 4, fig. 1).

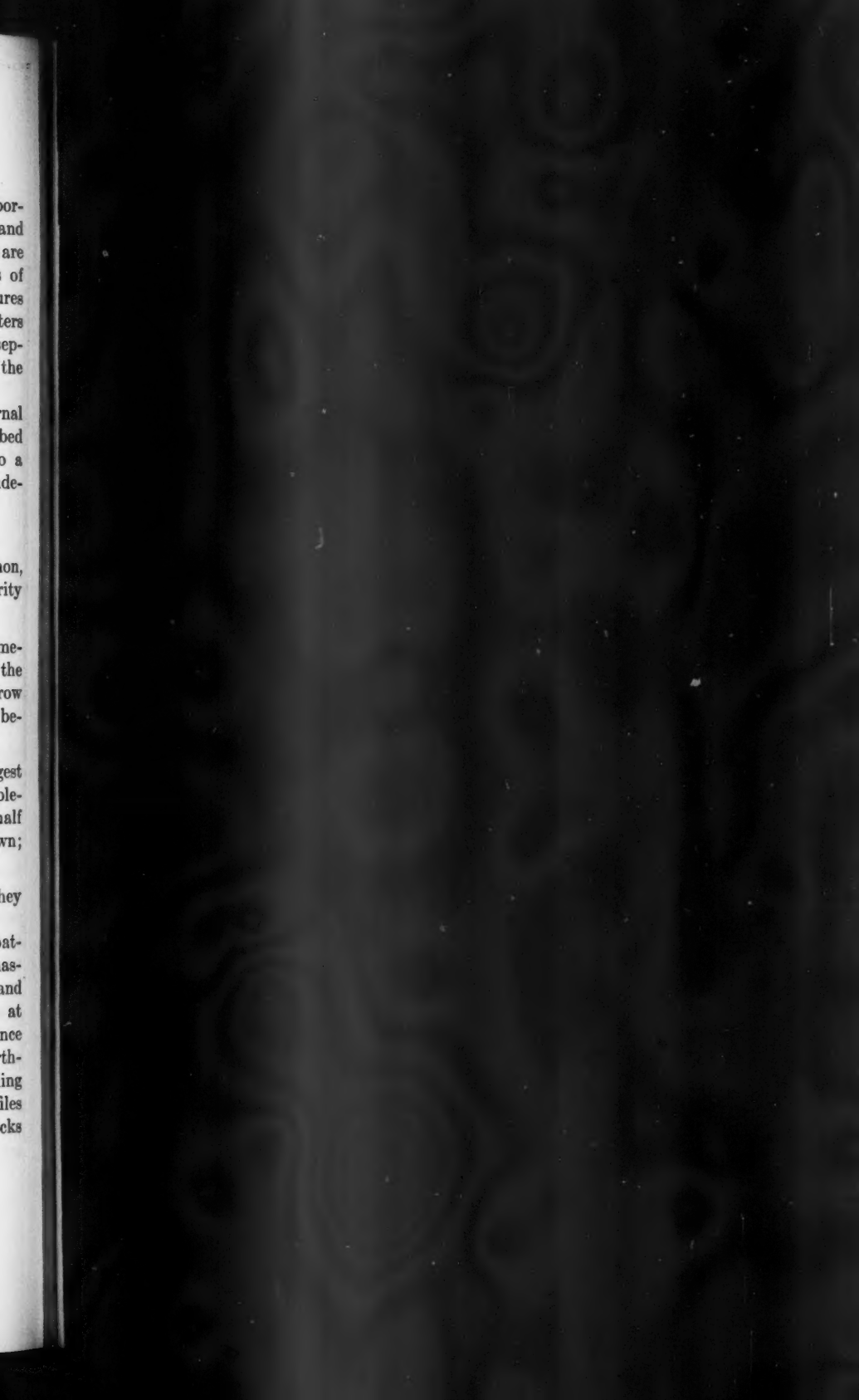


TABLE 2.—*Chart comparing features of the*

Plutons	Location and area exposed	Size and shape	Orientation	Type of pluton	Core a
Yankee Creek	*Secs. 26-35; T. 5 S. R. 72 W. Secs. 1-6, T. 4 S., R. 72 W. Secs. 25-27, 33-36, T. 5 S.; R. 73 W. Secs. 1-4, T. 6 S., R. 73 W., in parts. About 12 sq. mi.	9 mi. E. to W., 4 mi. N. to S. Sill of granite 1000 to 1500 ft. thick.	Sill intruded above quartz monsonite and below schist. Long axis of lensing sill is E-W.	*Uninuclear?* Few dikes of pegmatite or aplite, medium coarse trachytoid granite.	No well-defined magma spread W. and up thr openings clo at E.
Mason Creek	Secs. 15-17, 20-23, 29-30, T. 6 S., R. 71 W., in part. About 9 sq. mi. of granite and a little migmatite.	3 mi. E. to W., 3 mi. N. to S. Lenticular pod with thick edge to W.	Long axis, N. 60° W., lies along E. edge of Pikes Peak granite.	*Complex* Mass of fine-grained granite cut by pods of coarse granite; apophyses into Pikes Peak granite.	Three-centers o m. N. of S. ed of Schaffers en E. of Mason C
Turkey Ridge	Secs. 3, 4, 9, 10, and 16 of T. 6 S., R. 71 W. About 4 sq. mi., all granite; very little migmatite.	2½ mi. NW to SE, 2 mi. NE to SW. Asymmetrical—thick part to SW.	Asymmetrical lense thicker to the SW. Long axis N. 60° W.	*Uninuclear* One podlike mass with two parallel lobes.	N. of Turkey Turkey Ridge of action of co toid granite wi funneling out; tite or aplite.
Maxwell Creek	Secs. 5-8, T. 6 S., R. 71 W. Secs. 31, T. 5 S., R. 71 W. Secs. 36, T. 5 S., R. 72 W., in part. About 4 sq. mi. of granite.	2 by 2 miles across. Irregular podlike-fingering out to E. and W. into schist.	Long axis N. 10° W.	*Uninuclear* Less coarse trachytoid granite than other plutons; flow lines NS to N. 60° E.	One ill-defined action betwee Maxwell creek of granite NS. dip of flow la to NW.
Brookvale	Secs. 9-12, T. 5 S., R. 72 W.; also Sec. 15, in part. About 3 sq. mi., in three parts—Brookvale, Brookvale School, and Bald Mountain.	3 mi. EW by 1 mi. NS. Irregular pods of granite and sills.	Long axis nearly EW to N. 75° W.	*Complex* At E. is lenticular mass within schist; at W. a boss of granite in quartz monsonite with sills into schist at N.	Two centers of E. of Brookva at W. end of tain.
Brook Forest Evergreen Mountain	Secs. 17-21, 30-32, T. 5 S., R. 71 W. Secs. 24-25, T. 5 S., R. 72 W., in part. About 5 sq. mi. of granite lensing out to the NE and N.	2½ mi. NE-SW; 3 mi. NW-SE. Broad thick pod thickest near center of outcrop; at least 1500 ft. of granite exposed.	Long axis of lenticular mass lies N. 50-60° W.	*Complex* Similar to the Berrian-Doublehead pluton in general structure; a central complex core area flanked by migmatite and schist cut by lenses of granite.	Three closely areas—Harrya of Dunsinane Brook Forest directly N. of of feldspars, vertically folia bodies comma cores.
Riley Peak	Secs. 35 and 36, T. 6 S., R. 71 W., in part; about 1 sq. mi. of granite in three sub-parallel long narrow ridges of granite separated by schist dipping steeply NE.	Narrow but long subparallel NW-SE lenses of granite; about 1 mi. long by ½ mi. wide.	Strongly NW-SE axis, dipping steeply NE.	*Uninuclear* Subparallel lenses focusing from one central area; Dike festoons mostly oriented N. 60° W.	Mostly conceale big lensing po deploy from a s of action; othe lel lenses take central pod. end of central
Casto Creek	Secs. 31-33, T. 6 S., R. 70 W.; about 1 sq. mi. of granite in two long high ledges, on wall of Casto Creek valley.	From outcrop the granite body is about 1 mi. long by ¼ mi. thick; long lenticular body of several minor confluent parallel lenses.	Strike strongly N. 60° W.; dip of axial plane NE low angle.	*Uninuclear* Laccolithic pod with subparallel sill - like apophyses.	One center of act cupola masses through the cen pod of grani coarse trachyt foliated N. 6 fanned out flow lines to SE
Kinney-Marshdale	Secs. 13-15, 22-25, 36, T. 5 S., R. 71 W. Secs. 30-31, T. 5 S., R. 70 W.; about 3 sq. mi. of granite.	Tripart lobate outcrop, NW-SE trend of lobes and structures; about 3 mi. long by 2¼ mi. wide.	Long axis NW-SE; dip of lensing mass steeply to NE.	*Compound* Well-defined unit with cores from which granite spread lit-par-lit and laccolithic massive lenses.	Three centers of a lobe, E. of Ma center of SE. lo show swirls o funnels, and trachytoid text foliation NW. side of special
Berrian-Doublehead	Secs. 35, 36, T. 5 S., R. 71 W.; Secs. 1, 2, 11, 12, T. 6 S., R. 71 W.; Secs. 5-8, T. 5 S., R. 71 W.; about 5 sq. mi. of granite exposed.	Thick lobate massive body trending NW-SE to N-S; distinct western, central, and eastern parts; granite shows about 3 mi. NE-SW, and 2½ mi. NW-SE.	Long axis lies N-S to N. 40° W; axial plane dips steeply NE.	*Uninuclear* Well-defined unit deployed from one center of action; lit-par-lit injection into schist and gneiss at E. and W.	One large wel center of act Berrian Mounta foliation of coa toid granite, flow layers, fun and spires of co in fine migm "pepper and s borders.
Pleasant Park	Secs. 19, 20, 30, 31, T. 6 S., R. 70 W.; Secs. 24, 25, T. 6 S., R. 71 W.; about 2 sq. mi. of granite exposed over part of sections noted.	2 mi. EW by 1½ mi. NS; very irregular boundaries mostly concealed; shape roughly massive pod-like, of many closely spaced lenses and migmatite.	Axial plane and long axis strike N. 75° W., dip NE 30-40°?	Probably *Complex.*	One core area ic middle of Plea possibly anothe mi. NW. of it.
Independence-Bear Mountain	Secs. 1-3, 11-14, T. 5 S., R. 71 W., in part; about 5 sq. mi. of exposed granite in deeply lobate pattern.	2 mi. NE-SW and 4 mi. NW-SE; outcrops and structures elongated NW-SE; compound lenticular mass.	Marked NW - SE trend of axial plane; long axis lies N. 45-50° W.	*Compound*	Three centers of lobe of pluton, n of Independen tain, NW end Mountain belov
High Peak-Crow Hill	Secs. 16, 17, 21, T. 6 S., R. 70 W.; three small patches of granite—total less than ½ sq. mi.	Probably underlies most of High Peak and Crow Hill; shape seems to be irregular lobate; granite is mostly concealed by migmatite.	Appears to have long axis trending N. 20° W.	*Uninuclear?* Boundaries unknown.	Ill-defined, most under migmatit
Elephant-Butte Troutdale	Secs. 4, 5, 7, 8, 9, T. 5 S., R. 71 W., in part; about 3 sq. mi. of granite exposed.	3 mi. E-W by 1½ mi. N-S; egg shaped with big end to NE; general plan a big lenticular pod with cupola top and laccolithic branches.	Long axis trends N. 70° E. with dip of body steeply to NW.	*Complex* Nonlobate but festooned at SW. by many sills.	Three centers of ac end of Butte, c Troutdale, ½ m Troutdale, ma domical hills; l named shows feldspars, spire structures.
Hodgson School	Secs. 28-32, T. 5 S., R. 70 W.; about 1½ sq. mi. of granite exposed.	1½ mi. E-W by 1½ mi. N-S; massive lensing body with lobes at NW. and E.	Long axis nearly EW; dip ill defined for axial plane, appears to be NE low angle.	*Complex* Irregularly lobate marginal areas at N. and E.	Three centers of ac School 1 mi., SE NW of School mi; coarse gra and massive trachytoid grani
Legault	Secs. 7, 8, 17, 18, T. 6 S., R. 70 W.; Secs. 13, 14, T. 6 S., R. 71 W., in part; about 2 sq. mi. of granite exposed.	2½ mi. by 1? mi.; narrow N-S and elongated E-W; northern margin unknown.	Long axis EW; dip of axial plane N.	*Complex* Lobate at E., blunt at W.	Two centers of part of Sec. 13 tween Legault and Westwood feldspar masses hills.

* Locations given by section, range, and township may be found on the Denver Mountain Parks quadrangle of clear the space for geological features. Section locations are listed, however, to facilitate the reader's examinat

Core area	Roof	Walls	Floor	Special features
o well-defined core area; magma spread from E. to W. and up through several openings closely spaced at E.	Roof mostly eroded away, but originally the pluton was intruded under schist roof of Mt. Evans batholith and above the quarts monzonite.	Walls ill defined; at E. the granite lenses cut into the schist and gneiss E. of Yankee Creek; at W. granite eroded to vanishing edge.	Definite floor under the granite, and windows eroded through to show the quarts monzonite floor below.	Few observed; joints lie mostly N. 40-60° W. and dip NE; low angles.
hree-centers of action, ½ m. N. of S. edge, 1 mi. N. of Schaffers crossing, 1 mi. E. of Mason Creek.	Roof mostly eroded away, migmatite in place over core E. of Mason Creek, grades down into granite.	Sharp intrusive contact with Pikes Peak granite at W.; contact concealed at SE, confluent with Turkey Ridge granite at E.; fault border at S.	No floor noted; granite appears to continue at depth.	Thick E-W to N. 60° E. pegmatite dikes at E.; veins with auriferous pyrite, drusy quarts, and hematite crusts at SW.
. of Turkey Creek on Turkey Ridge, one center of action of coarse trachytoid granite with flow lines funneling out; no pegmatite or aplite.	No roof, no schist, no migmatite over the granite body.	Faulted against Mason Creek at S. blended with Maxwell Creek at NW, pinches out with schist of Blue Creek at N. and NE, and E.	No floor observed; granite appears to continue at depth.	Blocked in by faults at S. and E.; marked joints N. 60° W. dip NE, and N. 40-50° E. dip NW.
he ill-defined center of action between Cub and Maxwell creeks; foliation of granite NS. to N. 50° E.; dip of flow layers steeply to NW.	Granite fingers out into schist at W.; schist septum of Blue Creek rises over granite at the E. making part of the roof.	Faulted wall along Cub Creek at N dips to S low angle; intrusive contact at S. against Pikes Peak granite. Granite thins out lit-par-lit with schist at E. and W.	No floor noted; granite probably continues at depth.	Thick milky quarts dikes at W. close to schist edge; joints govern drainage pattern.
wo centers of action—one E. of Brookvale, the other at W. end of Bald Mountain.	Granite disappears under schist at NE and N.; migmatite above granite at W. and S., but no schist in place.	Sharp contact with walls at W. and SW of QM; granite here of dikes and small stock pattern; at N. and E. granite lenses out into schist roof, and walls unite.	No floor at W. but distinct floor of schist at N.; granite pinches out into schist at E. and rises abruptly against QM at S.	Ill-defined joints; granite has aplitic appearance on Bald Mountain; coarse granite only E. of Brookvale.
hree closely spaced core areas—Harrys Rocks, SW. of Dunsinane at N. of Brook Forest Hotel, and directly N. of Hotel; swirls of feldspars, funnels, and vertically foliated granite bodies common on the cores.	Granite mostly unroofed by erosion at W. and in central core area. At E. and NE, granite pinches out lit-par-lit and sill by sill into migmatite and schist of Evergreen Mountain.	Faulted wall against schist at W., faulted against granite of Maxwell Creek pluton at S., concealed wall at N., walls and roof unite at E. where granite lenses out into beds of schist, lit-par-lit.	Schist underlies granite at E. and NE but not at S., W., or N.	Marked N-S faults cross granite; joints on core area mostly NW-SE dip NE.; mineralized veins along W. edge.
ostly concealed core; one big lensing pod seems to deploy from a single center of action; other sub-parallel lenses take off from the central pod. Core at SE. end of central pod.	Probably the granite pods are cupolas of main granite body below now concealed; schist makes roof, which wraps the granite pods; little migmatite.	Walls, floor, and roof of schist enwrap the pods and lenses of this pluton; no migmatite selvage to granite.	Walls confluent with floor of schist which dips under granite lenses, the whole dipping to the NE about 40°.	None.
he center of action making cupola masses above and through the center of main pod of granite; mostly coarse trachytoid granite foliated N. 60° W. but fanned out radially by flow lines to SE.	Schist and gneiss beds curve over to top of massive granite pods and laccolithic lenses.	Walls, floor, and roof all one mass of schist which layer by layer enwraps the mass of the lenticular pluton body, pod by pod.	Floor of schist lies under lenses making up mass of pluton; mostly mica, gneiss, and biotite-sillimanite schist; floor as a whole dips NE about 50°.	None.
hree centers of action—NE. lobe, E. of Marshdale, in center of SE. lobe; all cores show swirls of feldspars, funnels, and spires of trachytoid texture; general foliation NW.—SE. outside of special structures.	Roof of schist and gneiss partly in place over middle of body; "tadpole" pods of granite lit-par-lit with schist of roof; no migmatite selvage to granite body; schist trends N. 40° W., dips steeply NE.	At S. schist faulted steeply against granite; at N. granite lenses down into schist beds of Little Cub Creek; at E. granite and schist walls concealed; at W. granite-schist contact follows foliation of schist in Lans Gulch-Turkey Creek.	Schist makes floor under NE lobe; no floor identified elsewhere under pluton.	Marked fault complex at S. end; strongly developed NW-SE fault series in middle of area, marked by drusy quarts and hematite crusts.
he large well-developed center of action E. of Berrian Mountain; marked foliation of coarse trachytoid granite, prominent flow layers, funnels, swirls, and spires of coarse granite in fine migmatite and "pepper and salt" granite borders.	Roof of migmatite and schist above granite; roof mostly eroded at W. in place at E.; large septum of hornblende gneiss E. of core probably part of roof structures.	Faulted ? schist—granite contact at S.; granite lenses down with foliation of schist at W. and N.; walls of migmatite and schist at E. confluent with roof.	No floor.	Pegmatite mostly on E. and W. flanks; none in middle; some adjacent to hornblende schist septum.
he core area identified in middle of Pleasant Park; possibly another about 1 mi. NW. of it.	At east, much migmatite between coarse granite below and fissile schist above; septae of schist at W. between lenses of granite.	Walls of pluton concealed under plowed fields at N.; granite lenses out at E. and W.; schist of roof continuous with that of walls, where granite lenses out.	No floor.	Sheared and fault zones NW-SE which thrust granite to SW in wedges with low angled fault sole.
ree centers of action, NW lobe of pluton, near summit of Independence Mountain, NW end of Bear Mountain below summit.	Patches of schist foliated NW–SE over middle of pluton; long NW–SE down-folded bodies of schist in granite at N.; thick cover of migmatite at SE; foliated NW–SE dip NE.	Roof and wall migmatite and schist continuous at E. and SE, foliation dipping steeply NE; at N. the EW lime silicate veins and bedded ridge make wall against granite; granite lenses out into schist folise of Little Cub Creek; granite faulted against granite at Stanley Park.	No floor.	Pronounced joint and shear planes trend NW-SE and dip NE.
-defined, mostly buried under migmatite.	Mostly in place; only on Crow Hill and High Peak do cupolas stand high enough to pierce the roof of schist and migmatite; flexures in schist and migmatite fit granite masses below.	Walls and roof not distinguished; they merge around the granite at E. and NE; steeply NE—dipping schist separates this pluton from the Legault pluton.	No floor observed.	None observed.
ree centers of action, at E. end of Butte, cliffs N. of Troutdale, ½ mi. south of Troutdale, marked by domical hills; last center named shows swirling feldspars, spires, funnel structures.	Sheath of migmatite over pluton at E.; roof mostly eroded at W. and S.	Well-defined granite-schist contact wall on N.; dips steeply N.; granite lenses out into schist at NE and E.; north limb of Wilmot Creek syncline makes S. wall; dip S. 30-40°.	Floor of schist under Troutdale center dips NE low angle; no floor observed elsewhere.	Joints and faults trend EW to N. 60° E. and dip steeply N. to NW.
ree centers of action, S. of School 1 mi., SE of school, NW of School about ½ mi; coarse granite lenses and massive bosses of trachytoid granite.	Roof mostly eroded away; rare patches over granite at N.	South wall of pluton faulted against mineralized vein zone; granite lenses out into schist beds at N. and NE.; E. end of pluton is buried in migmatite, which grades into schist at E.	Schist under granite at NW of pluton; dips to steeply NW; no floor elsewhere.	Mineralized fault area along S. border of pluton trends EW; dips N. at low angle.
vo centers of action, N. part of Sec. 13 and between Legault Mountain and Westwood; swirling feldspar masses on domical hills.	Roof probably in place at N. although outcrops are concealed; granite lenses out to E. and SE into schist.	Walls of Legault pluton are not visible.	No floor observed.	None observed.

s quadrangle of the U. S. Geological Survey. The grid is omitted from the maps of this paper to eader's examination of the areas in the field, since most of the area lies within 2 or 3 miles of a road.

The plutons stand out topographically. The granite cores are not the most conspicuous elevations; where the dikes and sills of the core have reinforced the wall rock of gneiss and schist, the riveted complex of granite and metamorphic rock withstands erosion best and now makes conspicuous mountains.

Orientation.—The plutons are concordant structures. They are accommodated to the foliation of enclosing schist and gneiss and to the massive outlines of older granite bodies.

The plutons have a long and a short axis and an axial plane. The axial plane carries the long axis of the lenticular granite body, which defines its strike and elongate structural pattern, and carries also the shorter axis (usually of intermediate length), which defines its dip (Fig. 3). In many plutons the long axis lies N. 10°-60° W., and the axial plane dips northeast 10 to 80 degrees. Several plutons lie athwart the general northwest—southeast alignment of the majority. The long diameter of the sill-like Yankee Creek pluton is east—west, and its axial plane dips N. 10°-20° (Pl. 2). The Elephant Butte-Troutdale pluton is elongated N. 60°-70° E., and the axial plane dips steeply to the northwest. The pluton south of Hodgson School trends east—west and dips steeply to the north. In general, the primary structural elements of the plutons are closely related to the long and short axial positions, and the topographic trends are closely patterned over the structures (Fig. 3).

Primary structures.—The plutons are discrete bodies developed from one or more centers where the magma deployed upward and outward into the adjacent schist and gneiss.

Structurally, the plutons are of three types: (1) Uninuclear—constructed about one center of magmatic spread; (2) complex—a fusion of granite streams from two or more closely spaced magmatic conduits shaped by the various forces which acted on their orientation during the intrusion, and (3) compound—coalesced and subparallel lensing bodies from several centers of action—"siamese twins"—of granite with peripherally merging lobes and lenses. Other plutons are stocklike and appear to spread at depth. Most of the plutons are elongated northwest—southeast.

Within the body of each pluton are primary structures which reveal its development (Balk, 1937); they may be seen on many outcrops. (1) In the coarser trachytoid granite of the lower levels of the pluton mass flow layers consist of tabular feldspar similarly oriented, with the longest axes of the crystals pointing in the direction of flow. Mica and quartz grains are accommodated to the interstices of the parallel feldspar laths. (2) The centers of action show several flow structure patterns:

(a) Wide swirls of feldspars strung out over an acre or more or confined locally to 6 or 8 square feet of granite surface (Fig. 4); (b) funnel patterns (which may be the three-dimensional form of the swirls) with the long axes of the feldspars oriented so that they almost mesh in funnel pattern; (c) vertically foliated acutely pyramidal bodies of coarse trachytoid granite cut upward through the fine-textured granite and migmatite and generally carry long tabular feldspars with the long axes in vertical position and the intermediate axes lying in flow planes striking N. 40°-60° W. in some plutons, and N. 60° E. in others. (3) Flow lines are the most easily discerned primary mineral patterns. Here, again, the feldspars are the clue to the magma movements. Mica crystals are small, scattered, and seldom clotted or strung out. (4) Inclusions are rare within the masses of the coarse trachytoid granite, and those which do occur line up parallel to the long axes of the feldspars (microcline). (5) "Ghost lines" in the migmatite surmounting the plutons and locally sheathing the main body of coarse granite indicate the behavior of the surrounding rocks during the emplacement of the plutons. (6) There are longitudinal, cross, and diagonal joint systems of magmatic origin. (7) Marginal joint patterns in the schist and gneiss surrounding the plutons follow closely the original foliation and bedding of the metasedimentary rocks, hence are not readily separated from the weathered partings which follow foliation. (8) Over centers of magmatic action the joint pattern is reticulated in east—west and north—south pattern, but the north—south-trending joints are the most pronounced.

Some plutons show vestigial roof structures; in others, the granite barely shows through the thick cover of migmatite and overlying schist. The contact of roofing schist and migmatite with the granite is rarely plain; cupolas and spinelike protrusions of granite into the metamorphic beds above make lumpy bosses which are mostly accommodated to the foliation and banding of the beds, but locally the granite masses crosscut the roof structures at low angles. Where the schist and gneiss of the roof make sharp contact against the granite of a massive pluton, a selvage of fine gray "pepper and salt" granite grades below into finely trachytoid granite with a few large feldspar crystals. The porphyritic granite grades downward into very coarsely trachytoid granite which shows in many places strong northeast—southwest foliation and platy flow structure. This coarse granite has no aplite or pegmatite dikes and carries no inclusions. Outcrops are massive and blocky.

In other places, sheared and drag-folded festoons of aplite and pegmatite dikes set in migmatite blanket the granite and lie between it and the schist and gneiss of the roof. The selvage rocks exhibit a sharp contact against the coarse trachytoid granite which is foliated northeast—

NORMAL FAULTS IN GRANITE OUTCROPS ON CUB CREEK ROAD

(C) is downthrown relative to B; and (B) is downthrown relative to (A). Ink lines show trace of fault plane; vegetation covers the drag; but grooved and slickensided granite lies at the roots.

FIGURE 1. TRACHYTOIDAL INDIAN CREEK GRANITE
At end of low ridge about 1 mile northeast of Evergreen. The long white crystals are microcline feldspar.

FIGURE 2. STREAKY APLITIC INDIAN CREEK GRANITE
Exposed on small granite ridge west of Hodgson School and north of the Highway along Turkey Creek, NW¼, Sec. 30, T. 5 S., R. 71 W.

TEXTURAL FEATURES OF INDIAN CREEK GRANITE

southwest. Flow structures in the mixed migmatite, aplite, and pegmatite selvage are planar and dip to the northeast at low angles. The coarse trachytoid granite below has a pronounced northeast—southwest lineation of the long tabular microcline crystals.

The plutons with the greatest thickness of roof in place show, locally, small podlike granite lenses close to the top of the pluton body, but detached from the granite below and in the schist and gneiss of the roof. The pods are stretched out subparallel to the contour of the granite below the roof, but their feeders are concealed (Fig. 13).

Plutons adjacent to the Pikes Peak granite are mostly bare; only vestigial gneiss and schist film the summit areas. They have commonly a simple structure, and each is oriented about a single center of deployment. The second and third rows of plutons are complex to compound and have a considerable part of the roof and selvage structure in place. The granite outcrops in the outer row of plutons do not permit definite determination of the structures below, but the plutons appear fairly simple in structure. At the north and northeast the flanks of the outer plutons are buried in migmatite; at the southeast the High Peak-Crow Hill pluton shows three bosses of granite which pierce the covering of metamorphic rocks. However, the domical pattern of migmatite streakings suggest a considerable bulk of granite directly below.

Beyond the outer belt of plutons only small dikes and sills of granite and aplite transgress the schist and gneiss. A fine-grained, almost structureless aplitic granite has converted the schist and gneiss locally into a pink quartzose and feldspathic rock. At the northeast, many flesh-colored clean-cut pegmatite dikes cross the regional northwest—southeast foliation trends at right angles.

A few plutons have been eroded so that parts of the floors are exposed. In places the bedded schists and gneisses of the Idaho Springs formation dip under granite lobes. In the same plutons schist and migmatite layers curve over the lumpy granite summits at the tops of the plutons. Even where the floors are well exposed, the conduits of the feeding channels from the source below into the plutons above do not show. The funnelling structures within the granite body, especially at the cores, suggest conduitlike connections to some reservoir area below.

Secondary structures.—Early doming of the region by the intrusion of the Rosalie lobe of the Pikes Peak batholith into the Idaho Springs metasedimentary beds imposed initial dips to the north, northeast, and east. Subsequent tilting of the region followed the lines of early pre-Cambrian uplift, and the plutons with their enclosing schist and gneiss beds were more steeply inclined to the east and northeast than in the

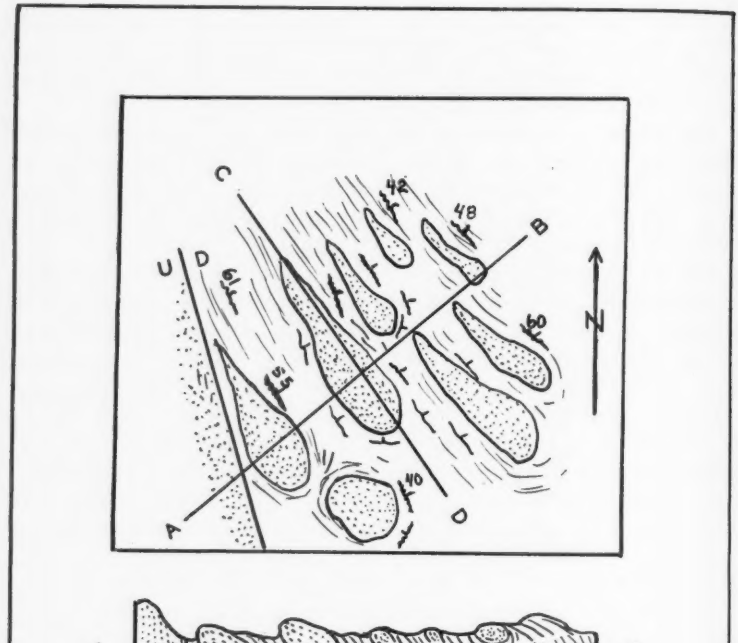


FIGURE 13.—*Sketch map and sections of the upper part of the Kinney-Marshdale pluton east of the center of action near Marshdale*

Stippled areas are well-defined outcrops of granite; intervening areas uncolored show by dip and strike symbols the foliation of the enclosing schist and gneiss.

first doming. Thus, although the interrelation of the plutons is much as it was during the pre-Cambrian, the inclination of the axial planes of the plutons in this part of the Front Range and to the east of the axis of uplift for this range is oversteepened.

The cross, longitudinal, and diagonal fracture systems of magmatic origin became the loci of movement and relief from tension and pressure during the strains incident to later movements in the Front Range (Pl. 8). The rectangular east—west and north—south vertical joint systems may have resulted from the faulting up of the plutons within their frame of schist and gneiss. Possibly, since the plutons are of lighter rock (granite) than the enclosing biotite-sillimanite-magnetite schist and gneiss, the strains incident to the adjustment of crystalline masses during the various uplifts of the Range since the pre-Cambrian, particularly the Laramide revolution, caused the plutons to rise above the heavier formations—i.e., to "pop up" when subjected to differential vertical strain.

Along Blue Creek low-angled thrust sheets of granite form three overlapping slices piled one above the other along the east side of the valley. The western edge of the Maxwell Creek pluton has been thrust to the west at a low angle over the Pikes Peak granite which outcrops at the eastern base of Black Mountain. Movement of the upper block has produced up-curving drag joints above the fault plane along which erosion has worked rapidly, producing the "bear caves" of local fame (Fig. 14). The soles of the low-angle thrust planes are striated and locally filmed with mica.

Nearly vertical faults trending N. 10°-40° W. have influenced the position of drainage lines across plutons and interpluton areas. Such faults have lowered some of the roof areas so that the schist and gneiss have been preserved over the granite. Many of the long vertical faults roughly block in the bodies of the plutons and are well developed south of Marshdale. There is no great mineralization of the long vertical faults. The faults which strike N. 10° W. have openings lined with drusy quartz and hematite crusts. The east—west-trending fracture zones have veins of green epidote, brown allanite, cinnamon-colored garnet, and milky quartz.

A series of fracture systems, N. 60° E., have been mapped which project from the margin of the Pikes Peak granite of the Rosalie lobe. These were probably formed soon after the intrusion of the Rosalie lobe, and their renewal during much later uplift—i.e., soon after the emplacement of the Indian Creek plutons—made way for extensive pegmatite and quartz veins (Pl. 3).

MODE OF INTRUSION OF THE PLUTONS

The Idaho Springs formation was intruded by three distinct major granitic bodies. The first great invasion brought the Mt. Evans quartz monzonite batholith and related smaller bodies into the region. This was followed, at no great interval, by the Rosalie lobe of the Pikes Peak

batholith. The final granitic invasion brought in the Indian Creek plutons.

The regional arrangement of the major crystalline units, intruded earlier in the pre-Cambrian, influenced the spread and localization of

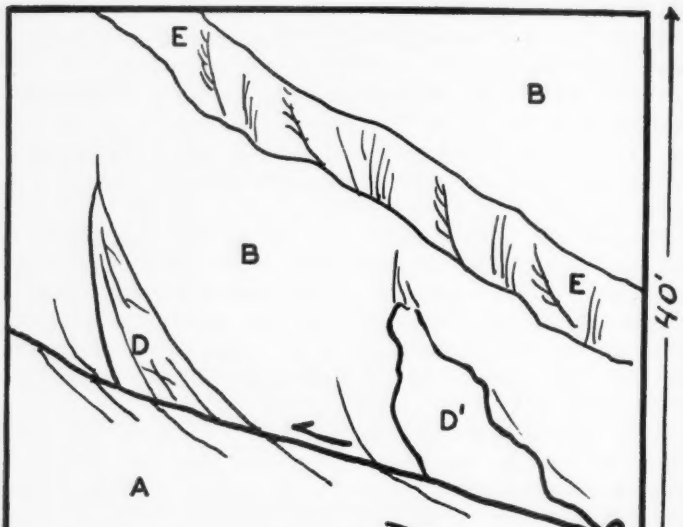


FIGURE 14.—*Thrust plane contact between the Pikes Peak and the Indian Creek granites on east flank of Black Mountain*

Sketched section of a vertical cliff facing south.

A. Pikes Peak Granite
B. Indian Creek granite
C. Thrust plane trace
D. Drag joints and sheared rock
D'. Cave at junction of drag joints and sole of fault
E. Pegmatite dike in Indian Creek granite

the Indian Creek plutons. At the south and southwest, the Pikes Peak batholith prevented ready escape of the fluid materials; at the west and northwest the batholith of Mt. Evans quartz monzonite hindered escape in that direction. But to the northeast and east, the foliation and fracture zones in the fissile Idaho Springs schist and gneiss offered easy egress to the Indian Creek magma. Hence, the plutons spread and developed in the only possible location—the re-entrant of Idaho Springs beds north and east of the Rosalie lobe of the Pikes Peak batholith.

The Mt. Evans intrusion created a rather simple dome on whose flanks were drag folds in the Idaho Springs beds. The stresses incident to the invasion of the Pikes Peak magma into the Rosalie lobe created marginal fractures and faults which affected both the Mt. Evans batholith

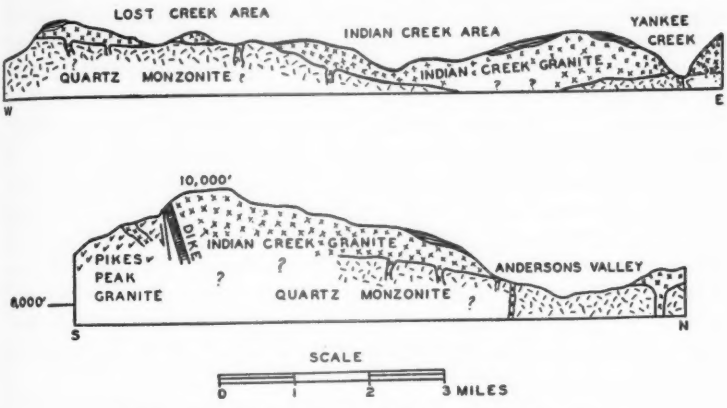


FIGURE 15.—*Sections across the Yankee Creek pluton area*

and the bedded structures of the Idaho Springs formation. These zones of weakness were channels which later controlled the movement of the Indian Creek magma. The structures in the Idaho Springs formation peripheral to the Mt. Evans batholith and to the Rosalie lobe of the Pikes Peak batholith controlled the emplacement of the Indian Creek plutons in rows along the curving margin of the Pikes Peak granite.

Within the coarsely crystalline Mt. Evans quartz monzonite batholith, breaks caused by the nearby intrusion of the Pikes Peak granite (Rosalie lobe) were few and irregularly spaced. At the summit of the Mt. Evans body there were sheet fractures at the junction of the quartz monzonite and its overlying roof of schist and gneiss (Idaho Springs). A long thin tongue of Indian Creek granite, thick at the east and thinning to the west, was intruded between the roof schist and the quartz monzonite. From the feeding channels through the quartz monzonite, the Indian Creek magma spread and banked up along the junction of the Rosalie lobe of the Pikes Peak batholith and the Mt. Evans batholith margin. Locally the Yankee Creek pluton has been eroded to the floor of quartz monzonite beneath it, and some of the pipelike channels which carried the Indian Creek magma up and allowed it to spread laterally above can be seen (Fig. 15).

The massive crystalline Pikes Peak granite in the Rosalie lobe resisted pervasive intrusion of the Indian Creek magma, but its weak peripheral marginal zone did not; and there, and along the fractures in the enclosing schist and gneiss, the sets of plutons were localized. The long axes of the Berrian-Doublehead, Kinney-Marshdale, Brook Forest-Evergreen and Independence plutons conform to the northwest—southeast trend of these fractures. The Hodgson School, Elephant Butte-Troutdale, and Brookvale plutons lie athwart the regional northwest—southeast strike and may have developed as the last of the pluton sequence, since they do not show the sheath of migmatite, below schist and above granite, which features those plutons elongated northwest—southeast.

The roughly semicircular foliation of the beds of Idaho Springs metasedimentary rocks, the drag folds, and tear faults developed by the riding up of the Idaho Springs beds on the flanks of the great compound anticline formed by the intrusion of the Mt. Evans batholith and the Rosalie lobe of Pikes Peak granite, and the marginal faults, shear zones, and cone sheet fractures within the Idaho Springs beds bordering the Rosalie lobe made a network of interlocking channels along which the upwelling Indian Creek magma traveled.

The advancing wave of Indian Creek mineralization made migmatite of the schist and gneiss along the main fracture zones, and replacement, plus infiltration lit-par-lit along the bedding and foliation of the metasedimentaries, created abundant aplite and small very fine-grained granite bodies.

Soon after the injection and spread of the Indian Creek magma, the Rosalie lobe of the Pikes Peak batholith began a series of collapses and failures within its frame of crystalline rocks. The fault planes along which the tension was relieved follow clearly the present outlines of the Rosalie lobe (Pl. 3). The subsidence of this semirigid crystalline body at its northern margin developed tension cracks slightly en echelon in the bordering rocks. The fractures were mostly N. 40°-60° E. in strike, and the breakage planes dipped generally to the northwest, parallel to and locally coinciding with the tear faults earlier developed on the eastern flank of the great anticline. The collapse of the Rosalie lobe also may have developed cone sheet fractures in the gneiss and schist which curved in semicircular pattern along the border of the Rosalie lobe on the east and north. The individual plutons were localized at the nodal positions where cone sheet fractures and radiating breaks intersected.

The Indian Creek magma within the magma chamber at depth had been largely stripped of mineralizing solutions, which escaped into the earlier fracture systems bordering the Rosalie lobe and in the Idaho

Springs beds. The residual magma, slushy with tabular feldspar crystals, deployed out into the schist and gneiss along the radiating fracture systems; collecting at these centers of action, it accumulated and spread into the schist and gneiss forming the plutons. The structures of plutons, already discussed, show a sequence of intrusion grading from aplite, pegmatite, and migmatitization into very coarse trachytoid granite. The initial escape of the Indian Creek mineralizers may have actuated the collapse of the Rosalie lobe nearby, and the failure of the Pikes Peak granite area may have induced further upwelling of the Indian Creek magma as it was expressed from the magma chamber. Well-defined fault zones bordering the Rosalie lobe of the Pikes Peak batholith have downthrown sides toward the Pikes Peak granite, indicating collapse of the lobe, at least locally.

The alternating collapse and upwelling apparently went on until most of the Indian Creek magma was used up in building the plutons. Each collapse of the Rosalie lobe made a network of fractures in the partly consolidated Indian Creek structures, and the dikes and sills, mostly with trends north—south to N. 60° W. and dips to the northeast of granite within granite and across it, are the fillings of the fractures. The plutons are, therefore, compounded of interlacing intrusions related magmatically to certain foci kept open as the magma repeatedly welled up and out.

The centers of action of most plutons are not symmetrically located within the body of the granite mass. The strike of flow lines and the pitch of the long feldspar axes indicate that the movement of the magma for the region as a whole was to the north and northwest and, deep within the plutons, to the northeast.

Each pluton developed individually as the amount of magma available permitted and as the stresses and strains of the governing shear and fracture planes and the zones of weakness invited. Some plutons were derived from a single center of action, others were compounded of several centers of action and their interfingering magma streams. Some are laccolithic and spread above a floor of schist and gneiss through one or more conduits, where they dilated into thick lenticular granite bodies between foliae of schist and gneiss; others are the granite fillings of drag folds on the flanks of the great domical structure centered by the Mt. Evans batholith and the Rosalie lobe of the Pikes Peak batholith.

Insofar as shown by structures related to the plutons, there was no "retreatal wave" of mineralization. Little or no pegmatite or aplite cuts the coarse granite of the cores or crosses the granite lenses which appear to be the last feature of magma spread. The wide, abundant pegmatite and aplite dikes which border the pluton areas originated mostly before the core structures were emplaced. Cooling and crystallization of the

last of the magma in the lower parts of the plutons developed shrinkage, causing longitudinal and cross joints and the sheeting and platy parting related to the flow layers.

The massive quartz veins which parallel the edge of the Rosalie lobe on the Indian Creek side of the contact constituted the final act of magmatic invasion. The last subsidence of the Rosalie lobe area broke the Indian Creek granite close to the contact, and the quartz-bearing fluids filled the fissures. Some of the Pikes Peak magmatic juices, either left over from the earlier crystallization of the Rosalie lobe or regenerated during the stresses and strains incident to the emplacement of the Indian Creek plutons nearby and the series of subsidences suffered by the Rosalie lobe, filled fractures in the Indian Creek granite and crystallized there, forming the pasty-appearing, red, aplite dikes characteristic of the Pikes Peak type of intrusion.

The Indian Creek plutons had been installed along the marginal fracture systems and fault zones which affected the Idaho Springs beds, converting the fissile and relatively incompetent beds into a strongly crystalline area, riveted and trussed by lensing plutons and their festoons of dikes and sills. Thus, the "neck" of the Rosalie lobe was encircled by an incomplete "necklace" of giant granite "beads" of Indian Creek granite.

In some ways the number, position, and alignment of the plutons around the border of the Pikes Peak granite mass resemble the structures in ring-dike and sill complexes described in the Tertiary of Ireland, Scotland, and some of the New England States (Anderson, 1924; Richey, 1932; Kingsley, 1935). The Pikes Peak granite of the Rosalie lobe subsided much as did the core structures of ring-dike complexes, and the plutons form rows of more or less adjacent bodies which, superficially at least, resemble the position and structures in ring-dikes related to cauldron subsidence. The granite, however, shows no differentiation from row to row of plutons, nor is there any progressive intrusion outward from the center. The Tertiary ring-dikes structures were not, apparently, intruded below any great overburden of crystalline rocks, as were the Indian Creek plutons. It is possible, although not probable, that the Indian Creek pluton set-up is a deep-seated pre-Cambrian pattern of much younger ring-dike and cauldron subsidence arrangements.

CONCLUSIONS

(1) The Indian Creek plutons are the youngest pre-Cambrian granite bodies of the area.

(2) The plutons were localized by fractures and folds in the Idaho Springs formation and were related to the intrusion of the earlier granite

masses of the Mt. Evans quartz monzonite batholith and the Rosalie lobe of the Pikes Peak batholith.

(3) The plutons were developed in a semicircular area peripheral to the north and east margin of the Rosalie lobe of the Pikes Peak batholith. They are not aggressive intrusions, but filled openings made by shear zones, or along the foliation of the Idaho Springs beds, or the cores of drag folds and spread out radially into the schist and gneiss along the border of the Rosalie lobe.

(4) There are at least 16 plutons, and each is an integral structure of podlike to lenticular pattern. Each pluton developed from one or more centers of magmatic deployment and is enwrapped by schist and gneiss.

(5) Within plutons, the flow lines, flow layers, primary joints, and fracture systems, together with inclusions and migmatite patterns, reveal the source of magmatic flow and its direction of spread from centers of action.

(6) The plutons are a fusion of magma streams which successively penetrated and interpenetrated, making intrusions within intrusions. Progressive subsidence of the Rosalie lobe of the Pikes Peak batholith made repeated upwellings of the Indian Creek magma, between which were quiet times of crystallization.

(7) The topographic expression of the plutons is directly related to the primary structures of the granite and to the interrelation of the plutons and the enclosing rocks.

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SOUTH SIDE OF MOUNT BOSWORTH

B = Bosworth, P = Pika, E = Eldon, S = Stephen, C = Cathedral, Pt-M = Ptarmigan? and Mount White, Sp = St. Piran formations, F = fault trace.

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LOWER AND MIDDLE CAMBRIAN STRATIGRAPHY OF SOUTHWESTERN ALBERTA AND SOUTHEASTERN BRITISH COLUMBIA

BY CHARLES DEISS

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ABSTRACT

The more important unsolved stratigraphic problems in the Waucobian and Albertan rocks of British Columbia in 1938 were: the geographic distribution and variation in thickness of the formations south and west of Ptarmigan and Castle Mountains; the origin and relationships of the Ptarmigan limestone to the Cathedral dolomite; the stratigraphic position of the Ross Lake (*Albertella*) shale on Mounts Bosworth and Assiniboine, and the *Ogygopsis* shale on Mount Stephen; and the position of the southern shores of the Cambrian sea on the north side of the Montana Island. The Mounts Bosworth and Assiniboine sections were remeasured and the Mount Stephen section was examined in 1939. The results are:

On Mount Bosworth the Ross Lake shale is well exposed 385 feet above the Ptarmigan limestone. The Eldon dolomite is 1110 instead of 2728 feet thick. The Lake Louise and Fort Mountain formations do not crop out, and only 750 instead of 2705 feet of St. Piran is exposed. The total thickness of the Mount Bosworth section is more nearly 6300 than 12,353 feet as originally reported.

The *Ogygopsis* shale seems to rest upon the Cathedral instead of beneath the Eldon on Mount Stephen, thus forming the lower instead of upper part of the Stephen formation. This problem remains unsolved.

The new Gog and Naiset formations are defined for the Waucobian and basal Albertan rocks in the Assiniboine section originally assigned respectively to the Fort Mountain, Lake Louise, and St. Piran, and Mount Whyte formations. The distribution and origin of the Ptarmigan formation are discussed.

Lower and Middle Cambrian geography and sedimentation are described, and the sections are correlated temporally with those in the Cordilleran area of the United States.

INTRODUCTION

THE PROBLEM

Since 1935 the writer has revised Lower and Middle Cambrian type sections and formations in the Cordilleran area (Deiss, 1936, 1939a, 1939b). The need for this revision and the problems involved were discussed in detail as each part of the work was completed. The field work in Canada was started in 1938 as the first step to obtain data for the

solution of the remaining broader problems of Lower and Middle Cambrian stratigraphy in the northern Cordilleran region. The problems which needed solution in 1938 (Deiss, 1939a, p. 954) were:

"(1) The published type sections in Alberta and British Columbia were not usable for accurate correlation, (2) the formation boundaries were drawn at different stratigraphic horizons in the different sections, (3) the exact stratigraphic horizons, and therefore the relative ages of many genera and species were unknown, (4) genera now known to be restricted to one of several widely separated stratigraphic horizons were lumped together in the published faunal lists, (5) some of the formations as defined were not mappable units, and (6) the Middle-Lower Cambrian boundary was incorrectly located."

These six problems were partly or entirely solved in 1938, but additional field work was needed to check adequately the conclusions reached and the interpretations of some of the stratigraphic relationships. The field work during 1938 also revealed seven new problems in connection with the Cambrian rocks of southwestern Alberta and southeastern British Columbia: (1) What is the geographic distribution and variation in thickness of the Ptarmigan limestone in the region south and west of the type section of the formation on Ptarmigan Peak, (2) was the Ptarmigan originally a separate sedimentary unit or does it represent the unaltered basal part of a much thicker limestone deposit of which the Cathedral formation is the dolomitized area, (3) what is the exact stratigraphic position of the *Albertella* zone on Mount Bosworth and Mount Assiniboine to the north and south of Ross Lake, (4) why is the *Ogygopsis* fauna and shale not present on Castle Mountain, Ptarmigan Peak, and Mount Bosworth only 6½ miles northeast of Mount Stephen, (5) what is the exact stratigraphic position of the *Ogygopsis* shale, and what are the temporal relationships of its fauna to the *Olenoides-Thomsonaspis* and *Glossopleura* zones of the Stephen formation on Castle Mountain, (6) what is the geographic distribution of the Pika formation to the northwest and southwest of the type locality of the formation on Castle Mountain, and (7) where were the southern shores of the Cambrian sea on the north side of the Montana Island?

In order to obtain field data bearing on these problems two sections were remeasured and a third was examined during the season of 1939: Mount Bosworth in Yoho National Park, Mount Assiniboine in southwestern British Columbia, and Mount Stephen in Yoho National Park. After completing the field work, the trilobites collected during 1938 and 1939 were compared with the types of described species in the United States National Museum. The results of the field and laboratory work constitute this report. The descriptions of the many new genera and species collected during the field work will require several years before they can be published.

LOCATION OF AREA

The sections studied are near the northwest and southeast corners of an area in southeastern British Columbia (Fig. 1) between [the parallels] 50° 53′ and 51° 30′ N. Lat. and [the meridians] 115° 35′ and 116° 27′

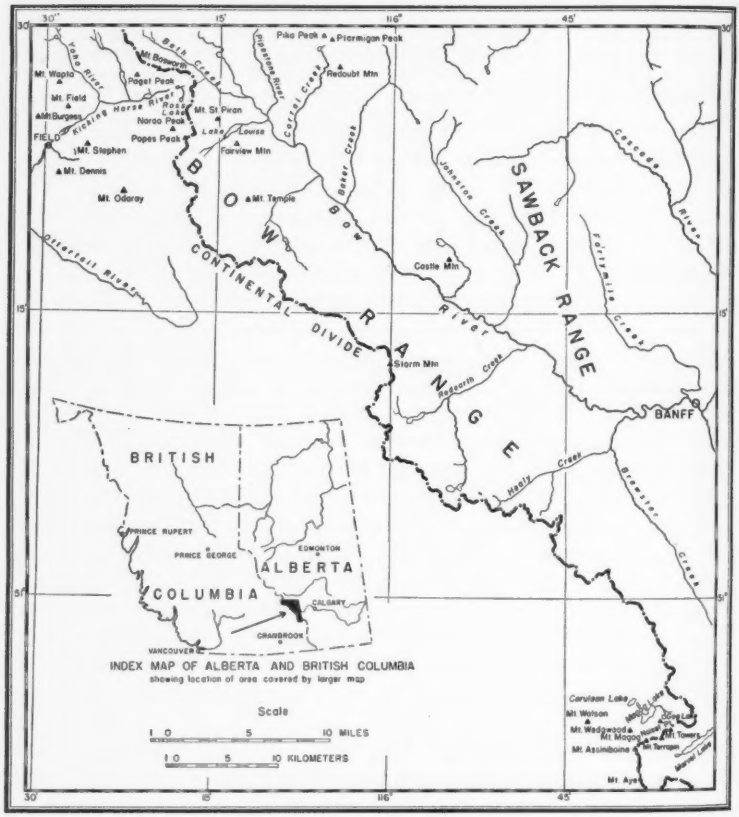

FIGURE 1.—*Map of area described*

W. Long. Mount Bosworth and Mount Stephen are in the northwest part of the area in Yoho National Park. Mount Bosworth is approximately 6½ miles N. 52° E. of Mount Stephen and 60 miles N. 36° W. of Mount Assiniboine. Both Mount Bosworth and Mount Assiniboine are parts of the Continental Divide and the Provincial Boundary between

Alberta and British Columbia. The sections, however, were measured in British Columbia. The Castle Mountain, Ptarmigan Peak, and Ross Lake sections measured in 1938 (Deiss, 1939a, p. 963-993) are in the northern half of the area and much closer to Mount Bosworth than to Mount Assiniboine.

ACKNOWLEDGMENTS

The field work was financed by a grant from the Penrose Bequest of the Geological Society of America. For this vital assistance the writer is deeply grateful and extends his thanks. Montana State University provided the field car, part of the equipment, all laboratory supplies and clerical services, and financed the writer's transportation to and expenses in Washington, D. C., thus making possible the work in the National Museum. To Dr. George Finlay Simmons, President of the University, the writer expresses his appreciation for this help. Major P. J. Jennings and Captain C. M. Russell, Superintendents of Banff and Yoho National Parks, gave much information concerning trails and camp sites, helped in obtaining maps, and granted special privilege to use "closed" National Park roads. Dr. C. E. Resser of the United States National Museum gave freely of his time and experience and made available to the writer the collections in the National Museum. Norman Denson assisted the writer in the field. His enthusiasm and interest added much to the successful completion of the field work. To each of these people the writer is indebted and extends his thanks.

SUMMARY OF PREVIOUS WORK

The published reports dealing with the Cambrian stratigraphy of the area were summarized, and a detailed chronologic account of the field work in this part of Alberta and British Columbia was given last year (Deiss, 1939a, p. 955-961). The writer is not aware of other field work in the area on these rocks and fossils, except that in 1939 upon which this report is based. Consequently, a restatement of previous work in the area is unnecessary.[1]

MOUNT BOSWORTH AREA, BRITISH COLUMBIA

GENERAL STATEMENT

The Mount Bosworth section has long been considered as "one of the largest and most complete Cambrian sections in the Canadian Rockies" (Walcott, 1928, p. 308-309) and has been cited and quoted in the Cam-

[1] In discussing the structure section published by Raymond and Willard and in the bibliography at the end of the paper (Deiss, 1939a, p. 960, and 1025) Dr. Bradford Willard's name was omitted. The publication describing the structure section was written jointly by Raymond and Willard (1931, p 97-116). The writer apologizes for the error and here corrects the mistake.

brian literature probably more often than any other section in western North America except that in the House Range of Utah. The Mount Bosworth section was first described by Walcott (1908b, p. 204-217) who assigned the upper 110 feet to the Ordovician and the lower 600 feet to the Fairview formation. In 1911 Allan and Burling discovered Upper Cambrian fossils in the rocks which form the upper 110 feet of Paget Peak (Walcott, 1912a, p. 229) thereby establishing the Upper Cambrian age of the beds. Four years later Walcott (1915, p. 178) summarized his Mount Bosworth section but omitted the upper 110 feet of beds which he previously assigned to the Ordovician, and again used the name Fairview for which three years earlier he had substituted the name Fort Mountain sandstone (Walcott, 1912b, p. 131, footnote a).

The lower part of the original section was much confused by the inclusion of part of the Lakes Louise and Agnes section located 5½ miles southeast of Mount Bosworth. The lowest beds which Walcott actually measured on Mount Bosworth were 503 feet of the St. Piran formation. This fact is clearly indicated by the following statement which he placed just below the description of the 503 feet of beds in his original section (Walcott, 1908b, p. 215-216):

> "The general dip of the strata is to the northwest 20°; strike north 30° east.
> "The section is continuous, with the exception of the displacement between the Paget and Bosworth formations of the Upper Cambrian. This does not cut out any considerable thickness of strata, as is proven by the unbroken section in the cliffs of Mount Daly three miles to the north.
> "In the Lakes Louise and Agnes section, about five miles southeast of Mount Bosworth, the total thickness of the St. Piran formation is 2,705 feet. Below the St. Piran the following section occurs:"

The last two sentences clearly state that the remainder and lower part of the Mount Bosworth published section was taken from the Lakes Louise and Agnes section where the Lake Louise shale is 105 feet thick and is underlain by 600 plus feet of "Fairview" (Fort Mountain) quartzitic sandstones. Thus Walcott created a composite section, failed to state the fact unambiguously, and then published the composite section under the heading "Mount Bosworth section." Confusion was added by the inclusion of the 2705 feet of St. Piran, 105 feet of Lake Louise, and 600 plus feet of "Fairview" in the résumé of the Mount Bosworth section which formed part of the same paper (Walcott, 1908b, p. 217) in which he had just referred these formations and thicknesses to the Lakes Louise and Agnes section. Then in 1915 Walcott (1915, p. 178) quoted a summary of his Mount Bosworth section in which he again gave the same formations and thicknesses taken from the Lakes Louise and Agnes section. Why this error and the resulting confusion and misinterpretation of the Lower Cambrian stratigraphy was not immediately corrected

is unknown, but it was only the addition of these beds from the Lakes Louise and Agnes section which increased the thickness of the Cambrian rocks in the Mount Bosworth section to "12,353 feet" and thereby made possible the statement that "a thickness of more than 12,000 feet (3,657.6 m) of conformable strata" (Walcott, 1917b, Pl. 1; 1928, Pl. 67) is exposed on Mount Bosworth and Paget Peak.

The problem was apparently only partly understood in 1928 when Resser published Walcott's sections (Walcott, 1928), but the exact conditions were not known, and the only other possible combination of errors was then made. The thickness of the St. Piran exposed in the Mount Bosworth section was stated to be 503 feet as originally given by Walcott; the Lake Louise shale was omitted from the Mount Bosworth section; but the 600 plus feet of Fort Mountain was retained. No reason or explanation for making these changes was given. In face of these facts (Walcott, 1928, p. 314), on Plate 67 of the same paper, the incorrect statement was again made that the photograph of the south side of Mount Bosworth "includes . . . a thickness of over 12,000 feet (3,657.6 m) of conformable strata." Thus by 1928 confusion concerning the lower part of the Mount Bosworth section was complete. These important changes in the original section (Walcott, 1908b, p. 204-217) were made in 1928 (Walcott, 1928, p. 308-314), and at the same time some of the errors in the original section were partly or wholly reproduced. Therefore, the emended section based upon the writer's field work in 1938 and 1939 is compared graphically with both the original section and that given in 1928 (Fig. 2). In order to discuss precisely the Mount Bosworth section which was already unnecessarily confused, Walcott's section published in 1908 is referred to as the "original" and the one published in 1928 as the "altered" Mount Bosworth section. (*See* discussion under heading "Comparison of original, altered, and emended Mount Bosworth sections.")

When the Mount Bosworth and all the other Cambrian sections Walcott measured in the Canadian Rockies (Walcott, 1928, p. 185-367) were published posthumously by Resser, several other important changes were made in the original section. (1) The upper 110 feet of beds was again omitted without explanation. (2) The lower 268 feet of the Bosworth formation was assigned (probably correctly) to the Arctomys formation. (3) Three units of the Cathedral "ld, le and lf" (Walcott, 1908b, p. 212) were omitted, thereby reducing the thickness of the formation from 1595 to 1212 feet. This change also was made without any reason or explanation. (4) The rocks originally assigned to the Cathedral were placed in the joint "Cathedral and Ptarmigan formations" (Walcott, 1928, p. 312).

Because the Bosworth section had been quoted so often by different workers and was reputed to be one of the most complete in the Canadian part of the Cordilleran trough, it should have been one of the most useful in the area as a standard for comparison. Unfortunately Walcott (1908b, p. 213) collected *Albertella* and other Middle Cambrian genera from a talus block and assigned them to the Lower Cambrian Mount Whyte formation. In 1915 Burling (1916, p. 469) discovered *Albertella in situ* in the Cathedral formation, 375 feet above the top of the Mount Whyte. The following year Walcott (1917a, p. 1-4) defined a new formation—the Ptarmigan—stated that it was present in the cirque wall above Ross Lake, and located the *Albertella* fauna in the section within the Ptarmigan formation. As the Ross Lake section is only 2½ miles southeast of the top of Mount Bosworth, the Ptarmigan formation should be present there also, but that section was not restudied by Walcott in 1917. In order to check these points before emending the section on Mount Bosworth, the writer remeasured the Ross Lake section in 1938 and accurately located and zoned the *Albertella* fauna in it (Deiss, 1939a, p. 989-992). As a result of this work several other errors appeared in the published Mount Bosworth section: (1) The Ptarmigan limestone was found to be absent at Ross Lake and from that point could not be observed on Mount Bosworth, (2) the Eldon dolomite seemed to be normal in thickness (1000 to 1200 feet) instead of 2728 feet as reported in the original section (Walcott, 1908b, p. 209), (3) the Stephen formation also seemed to be nearly normal in thickness (400 feet) instead of 640 feet as given, and (4) although the Arctomys formation was stated to rest on the Eldon (Walcott, 1928, p. 311) and the beds shown in this position on Plate 67 of the same paper are labeled "Bosworth", actually the rocks which form the upper part of Mount Bosworth (Pl. 1) appeared to belong to the Pika formation (Deiss, 1939a, p. 1008-1009).

These discrepancies and others known in 1939 necessitated remeasurement and careful restudy of at least the Lower and Middle Cambrian rocks on Mount Bosworth before that section could be used reliably as a standard for comparison or correlation.

LOCATION OF AREA

Mount Bosworth is part of the Continental Divide between Banff and Yoho National Parks and is in the northwest part of the area under consideration (Fig. 1). The peak lies at the south end of an unnamed range of mountains southwest of the Waputik Range and is several miles north of the Bow Range which terminates at Kicking Horse Pass. The strata on Mount Bosworth have an average strike of N. 17° W. and a dip of 20° SW. (Pl. 1) and form part of a monocline which is truncated on the

west by a high-angle thrust. Several small thrusts displace the beds locally on the mountain. The geologic structure on Mount Bosworth and the adjacent peaks to the west is similar to that in the north end of the Bow Range.

MOUNT BOSWORTH SECTION

The east end of Mount Bosworth is covered with dense timber, brush, glacial drift and soil to such an extent that a continuous section through the St. Piran sandstone is not exposed (Pl. 1). Consequently the emended section given here was measured from the lowest outcrop of continuous exposures on the southeast side of Mount Bosworth. The Fort Mountain quartzite crops out as isolated ledges in the timber between the base of the emended section and the valley floor of Kicking Horse Pass, but the outcrops are so separated by glacial drift that a continuous section cannot be measured (Pl. 1). The section traverse extends approximately N. 10° W. up the south slope of the mountain beginning at an elevation of 6000 feet above sea level, and N. 9° W. of Sink Lake and N. 80° W. of the intersection of the motor road and the summit of Kicking Horse Pass. The entire emended section was measured in the N. ½ secs. 3 and 4, and S. ½ secs. 8, 9, and 10, T. 29 N., R. 17 W.[2]

Neither the thickness nor the contact of the basal Cambrian quartzitic sandstone on the Beltian (Hector) shale is known on Mount Bosworth or at Ross Lake. Consequently the lower part of the Mount Bosworth section offers no evidence concerning the geographic extent or variation in thickness of the Lower Cambrian clastic formations northwest of Ptarmigan Peak.

UPPER CAMBRIAN

	Feet	Meters
Arctomys formation		
Shale and dolomite: buff and tan weathering; forms saddle on crest of ridge beneath Bosworth thick-bedded dolomitic gray limestone (estimated)	110	33.8
Dolomite: buff-tan weathering, thick-bedded, massive; forms ledge (estimated)	10	3.1
Dolomite and shale: buff and tan-gray, finely crystalline argillaceous dolomite in ½- to 4-inch beds, and some intercalated maroon and buff chunky shale. Forms buff tan slope above prominent red and buff ledge (estimated)	75	23.0
Shale: red and buff, finely arenaceous, chunky to fissile. Forms low sheer red cliff capped by 2 to 4 feet of buff shale on crest of ridge. Here top of measured section	15	4.6
Shale: as in underlying interval but nearly all red	31	9.5
Shale: as in underlying interval but here contains mud cracks and occasional bed of gray limestone fragments. Some chlorite and glauconite on a few beds	27	8.3
Shale: as in underlying interval but more buff and green and less maroon. Poorly developed glauconite on some upper beds	28	8.6

[2] The section numbers are based upon the Canadian system of numbering, which, unlike that used in the United States, begins in the lower right-hand corner of a township.

	Feet	Meters
Shale and limestone: dull-green, maroon and tan in alternate intervals of hard thinner-bedded shale and a few intercalated beds of salmon-buff argillaceous limestone spotted with dark-brown specks of limonite	26	8.0
Shale: deep-maroon and some thin intervals of green, fine-grained, hard, slightly sheared, thin-bedded, massive, finely arenaceous, maroon beds wavy banded with buff-tan clay. Weathers deep maroon stained buff	53	16.3
Shale: dark-maroon, buff, dull-gray, and very little green, hard, finely arenaceous, platy- and chunky-bedded (1/16 to 3/8 inch). Weathers maroon and buff in alternating intervals. Some thicker buff beds more or less dolomitic	40	12.3
Total thickness (measured and estimated) of Arctomys formation	415	127.5

Pika formation

Dolomite: pale-gray in lower 6 feet and pale-tan in upper 17 feet all finely crystalline, shaly- to thick- and smooth-bedded. Weathers pure buff and forms top of Pika formation here. Elevation approximately 8500 feet	23	7.1
Dolomite: light-dull-gray, finely crystalline, thick-bedded (6 to 18 inches), extremely massive and therefore appears unbedded in places. Some beds slightly petroliferous (fetid). Weathers light-gray stained dark-gray and to tiny sharp points on bedding and joint surfaces. A few intercalated beds of very light-gray, nearly lithographic, finely banded dolomite. Forms top of dark-gray band in upper part of Pika formation in this area	40	12.3
Dolomite: dull-gray, thick- and thin-bedded (⅛ to 2 feet), finely wavy-banded. Weathers light-gray and forms sheer cliffs	54	16.6
Dolomite: dull-gray, stained faint maroon on fresh fracture near weathered surface, nearly unbedded in lower 12 feet. Upper three-fourths pale-gray, extremely fine-grained, thick-bedded (½ to 4 feet), finely wavy-banded. Weathers light-gray and forms sheer cliffs	59	18.2
Dolomite: pale-lead-gray, very slightly vitreous, finely crystalline, thick-bedded (1 to 4 feet in lower part, 6 to 14 inches in upper part). Contains irregular masses (blebs and nodules) of lighter-gray, finely arenaceous argillaceous dolomite which weathers strongly in relief and to drab-tan color against dark-gray of dolomite matrix (Pl. 2, fig. 2). Upper 10 feet fine-grained, pale-gray dolomite which weathers dark-gray but contains no arenaceous clay nodules (estimated)	24	7.4
Interval covered on crest of ridge; in cliffs dolomite: buff-tan and gray weathering, massive, thick-bedded	22	6.8
Dolomite: dark- and light-gray, medium crystalline, thin- and platy-bedded (½ to 1 inch). Upper part black-gray, platy limestone which weathers black-gray and looks like isolated mass of limestone in dolomite here	25	7.7
Dolomite: as underlying, but generally thinner-bedded and more finely crystalline. Weathers to same light-gray color	36	11.1
Dolomite: black and light-gray mottled, medium to coarsely crystalline, thick-bedded. Weathers light-gray	33	10.1
Dolomite: black- and lead-gray, finely crystalline, very slightly vitreous, thick-bedded (½ to 4 feet), contains many veinlets of white calcite. Weathers dark-gray above buff-gray and is broken into large blocks shattered by ice here	32	9.8
Dolomite: gray and buff, thin- and thick-bedded, finely to medium crystalline, contains flakes, nodules, and blebs of tan clay. Weathers dull-gray and tan mottled, and is strongly mottled on fresh fracture	29	8.9

	Feet	Meters
Dolomite: white-gray, fine-grained, finely banded, thin-bedded (⅛ to ¼ inch). Weathers white-gray and to thin finely banded plates strewn over talus slope. Forms conspicuous marker here	1	0.3
Dolomite: dull-gray and faint-tan-white mottled, thick-bedded (8 to 18 inches), and some beds of dull-gray finely crystalline which contain buff clay flakes. Weathers lighter-gray and less buff than underlying dolomite	20	6.2
Dolomite: dull-gray, thick-bedded (2 to 10 inches), more or less strongly mottled. Contains irregular areas of clay. Weathers buff and gray mottled, and to vesicular cavities and sharp points and forms cliffs. Thicker-bedded upward	63	19.3
Dolomite: pale- and blue-gray, finely crystalline, banded lighter-gray in some beds. Weathers buff-banded dull-gray. A few beds of pale-blue-gray finely oölitic dolomite in lower part	34	10.4
Limestone: black- and dull-gray, finely to crypto-crystalline. At base 4-foot bed, generally medium bedded (½ to 6 inches, average 2½ inches). Some beds finely oölitic, others contain blebs and flakes of calcite and calcareous finely arenaceous clay. In lower half several 5- to 12-inch intervals of shaly-bedded drab-tan limestone and finely arenaceous chunky shale	51	15.7
Limestone: same dolomitic limestone as in underlying interval but thicker-bedded. Weathers pale-gray	8	2.5
Limestone: light- and dull-gray, finely crystalline, dolomitic, platy- to thick-bedded (½ to 8 inches) in alternate intervals, finely banded, hard. Weathers pale-gray in contrast to dark-gray underlying interval	24	7.4
Limestone: dark-gray, thin-bedded in lower part, thick-bedded (4 to 20 inches) in upper part. Weathers dark-gray and to platy fragments	37	11.4
Limestone: black-gray, fine-grained, thin-bedded, massive. Weathers dark-gray and to sharp points on joint and bedding surfaces, and breaks down to angular fragments	21	6.5
Limestone: dull-gray, shaly-bedded and mud-cracked in lower 2 feet, becoming dark-gray, oölitic and thicker-bedded upward. Weathers dark-gray	7	2.2
Limestone: dark-gray, finely crystalline, thick-bedded, finely banded. Contains excessive numbers of tiny pyrite crystals. Limestone weathers pale-buff in lower third, and bright-tan in upper two-thirds. Color of weathering results from alteration of pyrite to limonite. Forms another conspicuous buff-tan band in cliffs	5	1.5
Limestone: dark- and black-gray, finely to crypto-crystalline, thick- and thin-bedded. Contains irregular amounts of clay flakes and nodules. Weathers dark-gray and forms cliffs	14	4.3
Limestone: pale-gray, nearly lithographic, finely banded lighter gray. Weathers pale-buff-tan	3	0.9
Limestone: dark-gray, fine-grained, shaly-bedded in lower and thick-bedded (3 to 8 inches) in upper part. Some beds wavy banded. Weathers dark-gray and to sharp points on thicker beds	12	3.7
Limestone: one bed, bright-orange-tan and buff-brown-weathering, pale gray on fresh fracture, fine-grained, argillaceous. Breaks to conchoidal angular fragments and forms conspicuous marker in cliffs here. Contains tiny crystals of pyrite as in upper 5-foot buff interval	1.5	0.5
Limestone: dull-blue-gray, fine-grained, wavy bedded. Contains many irregular stringers and flakes of drab-tan clay. All much sheared here	1.5	0.5
Limestone and shale: light- and dark-gray, hard, more or less platy, fine-grained limestone finely and irregularly banded lighter-gray, and intercalated 2- to 4-inch beds of dark-gray (black) finely crystalline, hard, smooth-bedded limestone. Some ½- to 1-inch beds of dull-green arenaceous shale	23	7.1

	Feet	Meters
Limestone: dull- to dark-gray, generally fine-grained, thin-bedded in lower part, thicker in upper part (1 to 10 inches) and more regular. Contains much pale-tan finely arenaceous clay as flakes and nodules, and occasional worm borings. Weathers black-gray and forms cliffs. This is basal member of lower black-gray band in Pika formation in this part of Yoho National Park	45	13.8
Limestone and shale: dark- and dull-gray, fine-grained, argillaceous, thin- and irregularly shaly-bedded limestone, interbedded with tan arenaceous clay as thin irregular beds, partings, flakes, and nodules. Beds much cleaved locally. Weathers to tan slopes, and on cliff-faces beds banded tan and gray. Forms upper part of buff zone at base of Pika on Mount Bosworth	39	12.0
Interval covered: float and soil and glacial drift. Contact of Pika on Eldon covered. May be an error of 10 to 40 feet in thickness in section at this point	58	17.8
Total thickness of Pika formation	845	260.0
Total thickness (measured and estimated) of Upper Cambrian	1260	387.5

Note: The section was traced S. 55° W. to the ridge between Mount Bosworth and Paget Peak and continued up the ridge to an elevation of 8400 feet. The section was traced and correlated on the Pika-Eldon contact.

MIDDLE CAMBRIAN

Eldon dolomite

	Feet	Meters
Dolomite: white-buff-gray, medium-coarsely crystalline, thick- and thin-bedded (1 to 3 feet), massive. Weathers tan and buff-gray (estimated)	150	46.2
Dolomite: dull- and pale-tan-gray, coarsely crystalline, thick-bedded, slightly banded black-gray, slightly mottled. Contains stringers of lead-gray dolomite in some beds. Interval covered with talus	59	18.1
Dolomite: tan-buff-white, coarsely crystalline, irregularly banded dark- and light-gray, thick-bedded. Weathers dull-gray stained buff in large irregular areas	48	14.7
Dolomite: white-gray, coarsely crystalline, thick- and thin-bedded (8 to 24 inches). Some beds mottled light- and dark-gray, others with red-brown stain near surface on fresh fracture. Weathers dull-gray spotted buff	61	18.8
Dolomite: as underlying interbedded with intervals of thin-bedded dolomite (2 to 4 inches) and is lead-gray in upper few feet. Beds 2 to 8 feet thick. Weathers pale-gray	68	20.9
Dolomite: pale-steel-gray, finely to medium crystalline, much thinner-bedded (1 to 3 feet) than underlying. Contains blebs of buff-tan-weathering calcite. Weathers dull-gray, some beds stained dark-gray. Forms broken ledges and large angular talus blocks	97	29.8
Dolomite: faint-buff-white-gray, massive and thick-bedded as in underlying interval but much more coarsely crystalline. Weathers faint-buff-gray and forms part of some sheer cliffs	37	11.8
Dolomite: white-gray, medium crystalline in 8- to 20-foot beds, appears nearly unbedded. Weathers faint-buff-gray and forms part of massive cliffs with underlying dolomite	58	17.8
Dolomite: light- and dark-gray, thick-bedded as in underlying interval. Weathers light-buff-gray with large irregular areas		

	Feet	Meters
stained dark-gray. Very sugary and slightly mottled on fresh fracture. Forms broken cliffs. May be error of 5 to 10 feet in this thickness as beds could not be accurately traced around the shoulder of an unclimbable cliff	50	15.4
Dolomite: as underlying, but near base of interval is 10 to 12 feet of thin-bedded (4 to 8 inches) bright-tan-weathering dolomite	47	14.4
Dolomite: white-gray, medium to coarsely crystalline, thick-bedded. Weathers white- and faint-buff-gray and to vesicular cavities and to rounded cliffs and boulders	45	13.8
Dolomite: dark- and light-gray, and some buff, finely to medium crystalline, more or less thin- and thick-bedded. Weathers dark-gray mottled light-gray in lower 3 feet and buff in upper 5 feet, and to vesicular cavities. One 2- to 3-inch bed of coarsely crystalline white buff calcite	8	2.5
Dolomite: light-buff-gray, coarsely crystalline, almost unbedded, massive, light-gray and faint-buff on joint planes	15	4.6
Dolomite: light- and dull-gray, medium crystalline, more or less banded and thinner bedded in lower part. Weathers faint-buff and light-gray and forms broken cliffs	30	9.2
Dolomite: white-buff-gray, medium crystalline and some intercalated beds of coarsely crystalline, in ½- to 6-foot beds, extremely massive. Weathers pale- and buff-gray and forms sheer cliffs	58	17.8
Dolomite: pale-blue-steel-gray, finely to coarsely crystalline, extremely thick-bedded and massive, mottled with nodules of white crystalline calcite. Weathers buff-gray and to long vesicular cavities on joint planes and to rounded surfaces	41	12.6
Dolomite: faint-tan-gray, finely crystalline, massive. Weathers buff-tan-gray	10	3.1
Dolomite: as underlying but thicker-bedded, more massive, and contains blebs and irregular stringers of buff-weathering clay. Weathers light-gray and forms cliffs. Lower beds mottled light- and lead-gray. Dark areas weather very slightly in relief	64	19.7
Dolomite: light- and dull-gray, fine-grained, slightly argillaceous, hard, thin-bedded (1 to 5 inches), some beds mottled light- and dull-gray. Weathers pale-gray and forms broken cliffs	39	12.0
Dolomite: pale- and buff-gray, medium crystalline, hard, thick-bedded (2 to 8 feet), massive. Weathers light-gray mottled dark-gray and to vesicular cavities ½ to 2 inches in diameter. Upper part thinner-bedded, more coarsely crystalline, and darker-gray	40	12.3
Interval covered here: underlain by light-gray-weathering, finely crystalline dolomite in 2- to 8-foot beds. Forms cliffs to west of section traverse	44	13.5
Dolomite: dark-gray, banded irregularly pale-gray, finely crystalline, dense, hard, irregularly bedded (3 to 36 inches). Weathers dull- and pale-gray banded in ½- to 1-inch bands	19	5.8
Dolomite: tan-gray mottled with stringers of lead-gray, crystalline, thick- and thin-bedded (2 to 12 inches). Weathers tan-buff-gray, mottled in irregular gray areas. Forms broken cliffs and base of Eldon dolomite here	22	6.8

Note: The section was traced west six-tenths of a mile to a point approximately 6700 feet in elevation N. 28° E. of Wapta Lake Lodge. The section was traced and "tied in" on the contact of the upper black-gray thin-bedded Stephen limestone against the light-buff-gray medium-bedded buff-weathering basal Eldon dolomite. The upper 150-foot dark limestone interval of the Stephen thickens westward to here, whereas the underlying shale intervals seem to thin as though

	Feet	Meters
the shale were replaced by limestone in the upper part. Also, the upper black-gray limestone is much thicker-bedded and more massive at this locality than it is six-tenths of a mile to the east where the Stephen section was measured, and is transitional upward into the overlying dolomite here.		
Total thickness of Eldon dolomite	1110	341.6
Stephen formation		
Limestone: black-gray, fine-grained, in 1- to 6-inch beds, hard, irregularly bedded. Weathers dark-gray and forms upper part of dark-gray cliffs of Stephen formation here. Elevation approximately 7700 feet (Barometer)	48	14.8
Limestone: sooty-black, finely crystalline, hard, thin-bedded (1 to 3 inches). Weathers black-gray and forms sheer cliffs. Contains blebs, veinlets and stringers of white calcite, and occasional flakes and nodules of tan- and salmon-buff clay. In gulch bottom an isolated mass approximately 100 feet in diameter of pale-gray dolomite slightly mottled dark-gray exhibits same bedding as surrounding black-gray limestone	39	12.0
Note: Section now traced west 600 feet to bottom of next gulch and continued there to the base of the Eldon dolomite.		
Limestone: black-gray, fine-grained, hard, thin- and irregular-bedded. Similar to underlying. Weathers black-gray and forms steep slope and lower part of dark cliffs	45	13.8
Limestone: dark- and black-gray, fine-grained, thin-bedded (½ to 2 inches). Contains brachiopod and trilobite fragments. Weathers dark-gray and forms base of dark-gray cliffs	19	5.8
Shale, limestone, and conglomerate: gray-green and faint-tan, fissile and chunky metamorphosed shale, 6 to 8 feet of dull-gray, thin-bedded buff-weathering limestone in one unit, and several beds of intraformational conglomerate intercalated in shale below limestone. Weathers tan-gray and forms top of upper slope below dark-gray cliffs of Stephen formation on south side of Mount Bosworth	28	8.6
Shale and conglomerate: pale-gray and pale-green-gray, argillitic, hard, metamorphosed shale. Contains some talc and sericite and is finely arenaceous in some beds. Weathers green and tan. Several intercalated 4- to 15-inch beds of poorly developed intraformational conglomerate. Matrix dove-gray, fine-grained, and contains many blebs and flakes of buff-brown clay and some glauconite. Pebbles of faint-tan-gray fine-grained limestone. Interval forms slopes	48	14.8
Shale and limestone: green and greenish-gray calcareous metamorphosed hard chunky shale and some interbedded dull- and dark-gray massive but shaly limestone. Weathers tan and forms slopes and broken cliffs. Some shale finely arenaceous. Middle part of interval forms broken cliffs between upper and lower shale slopes of Stephen formation on south side of Mount Bosworth	64	19.7
Interval covered: float tan and green argillaceous shaly limestone and green calcareous shale: Upper part blue-gray, argillaceous, fine-grained, thin-bedded limestone intercalated in rusty-tan arenaceous shale, all metamorphosed and partly cleaved. Forms lower slope above black-gray cliffs	45	13.8
Limestone: black-gray, fine- to coarse-grained, thick-bedded as underlying interval. Contains stringers and specks of bright orange-tan limonitic clay and irregular oval areas which simulate *Girvanella*	5	1.5

	Feet	Meters
Limestone: dull-gray, fine- to medium-grained in 1- to 4-foot beds. Weathers dark-gray	8	2.5
Limestone: as in underlying interval but more black-gray and thicker-bedded. Near middle a 2- to 3-foot interval of black-gray fine-grained smooth bedded limestone finely banded white-gray. Much wavy banding throughout	26	8.0
Limestone: dark-gray, fine-grained, thin-bedded (2 to 6 inches, average 3 inches), hard, flaggy bedded. Contains varying of tan clay. Weathers tan-gray on slopes, but forms dark-gray cliffs	59	18.1
Limestone: dark blue- to lead-gray, fine- to medium-grained, fossiliferous, interbedded with tan-gray, argillaceous fine-grained platy limestone which contains much clay as partings, and many worm-tubes on some beds. Occasional 1- to 6-inch beds of blue-gray limestone. Fossil loc. C7a in basal part of interval. Typical *Glossopleura boccar* zone, but fossils relatively rarer than in this zone elsewhere	30	9.2
FOSSILS: *Glossopleura* cf. *boccar* (Walcott) *G. fordensis?* Deiss *G. nitida* Resser		
Limestone: dark-blue-gray, fine-grained, argillaceous, thin-bedded (¾ to 4 inches). Contains wavy bands of dull-tan calcareous clay, which weather bright-tan and in relief on joint faces. A few ⅛- to ⅜-inch masses of pyrite in lower beds	6	1.8
Note: The section was traced west and "tied in" on the contact of the Stephen-Eldon formations. The above part of the section was measured on the divide ridge between the third and fourth streams west of the east end of Mount Bosworth at a point approximately 7300 feet in elevation and N. 36° W. of the west end of Sink Lake.		
Total thickness of Stephen formation	470	145.1
Gouge: tan and gray, fissile, sericitic, contains finely disseminated talc. Marks fault which here throws upper part of Stephen limestone against upper part of Cathedral dolomite. West side downthrown	1–2	0.3–0.6
Note: Here a fault cuts most of the Stephen formation and a little of the upper Cathedral on the crest of Mount Bosworth (*see* F on Plate 1) at an elevation 8250 feet. The contact of the basal Eldon dolomite on the black Stephen limestone is clearly exposed 200 to 250 feet west of the fault here.		
Cathedral dolomite (type locality)		
Dolomite: buff- and some dull-gray, slightly mottled, thin-bedded. Weathers buff and buff-tan. Much distorted by ice and faulting here. Forms highest point (stratigraphically) measurable in Cathedral on this part of Mount Bosworth. May be error of 25 to 50 feet of thickness in upper part of Cathedral dolomite (estimated)	50	15.4
Dolomite: light- and dark-gray mottled, medium crystalline, in 8- to 15-inch beds. Weathers buff-tan and forms small peak with monument on top	57	17.5
Dolomite: as underlying interval, but less mottled and weathers to yellower-buff color	38	11.7
Dolomite: buff, some mottled gray, more or less thin-bedded, thicker-bedded up section. Occasional beds of dark- and pale-gray mottled. Weathers buff and to small fragments and forms part of flat crest of ridge here	184	56.6

	Feet	Meters
Dolomite: pale-buff, mottled dark-gray, medium crystalline, in 3- to 18-inch beds. Contains large masses and blebs of calcite. Weathers buff-tan mottled gray	63	19.4
Limestone: black-gray, fine-grained, thinner-bedded than underlying 12-foot interval. Weathers dark-gray and to small angular fragments, and forms bench on crest of east spur of Mount Bosworth	7	2.2
Dolomite: pale-buff, medium crystalline, thick-bedded (1 to 4 feet). Some beds mottled dark-gray. Weathers bright-tan	12	3.7
Limestone: black- and dark-gray, fine-grained, slightly argillaceous, in 1- to 7-inch beds. Contains veinlets of white calcite and blebs of gray clay. Weathers dark-gray	22	6.8
Dolomite: buff-gray, finer-grained than underlying, thin-bedded. Weathers pale-buff and to small fragments. Much covered with talus here	39	12.0
Dolomite: thin-bedded. Weathers buff. One 8-foot interval mottled black and dull-gray, and weathers dark-gray	18	5.5
Dolomite: white-buff, coarsely crystalline, some beds mottled dark-gray. Thick- and thin-bedded (⅓ to 3 feet). Weathers buff and buff-tan	36	11.1
Dolomite: pale-buff and white-gray, medium to coarsely crystalline, in 4- to 8-inch beds. Weathers bright buff and some salmon buff	32	9.8
Limestone: pale-dove-gray in upper three-fourths, fine-grained, hard, thick- and thin-bedded. Contains many blebs and stringers of white calcite which weather buff and in relief to sharp points. Limestone weathers pale-gray. Here contact of dark-gray limestone and bright-buff dolomite near east crest of spur of Mount Bosworth	39	12.0
Limestone: as in underlying interval, but more massive and thicker-bedded. Contains stringers and blebs of white calcite. Weathers dark-gray	22	6.8
Limestone: dull-black-gray, fine-grained, pure, argillaceous, in ¼- to 5-foot beds, massive. Contains flakes and nodules of gray and drab-tan clay. Weathers dark-gray and to angular fragments and forms rounded cliffs	53	16.3
Limestone: black- and dull-lead-gray, pure, fine-grained, argillaceous, thin-bedded (1 to 4 inches, average 2 inches), massive. Contains tiny blebs and specks of orange-tan limonitic clay and many irregular stringers of finely arenaceous clay which weather drab-brown and in relief. Limestone weathers dark-gray and to angular fragments, and forms base of cliff here	13	4.0
Limestone: dark-blue-gray to black-lead-gray, fine-grained, argillaceous, thin-bedded, much sheared. Contains dull-tan clay flakes and nodules which weather orange-tan and buff. Forms undercut sheared zone in cliff just above Ross Lake shale	5.5	1.7
Ross Lake shale member		
Shale and limestone: dull-gray, calcareous, unfossiliferous, irregularly bedded shale which weathers drab-gray and to conchoidal fragments, and a few intercalated lenticular beds of gray fine-grained limestone. Forms top of Ross Lake shale member, but is generally covered in this section	1.5	0.5
Shale: dull-gray, becomes more green and tan and thinner-bedded upward. Weathers drab-green and brown and contains numerous specimens of *Albertella* fauna as in lower 2-foot interval	5.5–5.7	1.7–1.8
Limestone: one bed: dull-gray, fine-grained. Contains a few blebs and flakes of gray and tan clay. Bedding surfaces irregular. Unfossiliferous here, but contains *Ptarmigania* at Ross Lake	0.2–0.3	0.06–0.1

	Feet	Meters
Shale: dull-gray, hard, more or less siliceous, platy, smooth-bedded, extremely fossiliferous, very slightly calcareous. Weathers to plates stained tan-brown and drab-green. Contains same species as upper 5.7-foot shale interval above. Entire Ross Lake shale zone is Fossil loc. C6e	1.8–2	0.5–0.6
FOSSILS: *Albertella bosworthi* Walcott; *A. helena* Walcott; *A. nitida* Resser; *A. robsonensis* Resser; *A. rossensis* Resser; *A. similaris* Resser; *A.* n. sp.; *"Amecephalus" stator* (Walcott); *"A."* n. sp.; *Kochina bosworthensis* Resser; *Ptarmigania rossensis*? Resser[3]; *Vanuxemella nortia* Walcott; *"Acrothele" colleni*? Walcott; *"Wimanella" simplex* Walcott; *Hyolithes cercops*? Walcott		
Total thickness of Ross Lake shale member	9.5	3.0
Limestone: black-gray in lower and paler-gray in upper part, fine-grained, thin-bedded, hard in lower third, massive and contains many clay flakes and partings in upper two-thirds. Generally covered and forms bench below Ross Lake shale	10	3.1
Dolomite: pale-tan-gray, mottled dark-gray, coarsely crystalline, thick-bedded (1 to 4 feet). Weathers drab-buff and forms cliffs here. Much jointed and broken down to angular fragments. Forms top of buff cliffs and slopes below Ross Lake shale here	37	11.4
Dolomite: white-buff, coarsely crystalline, extremely massive and thick-bedded (2 to 7 feet). Weathers buff-tan	38	13.6
Limestone: black-gray, dolomitic, thick-bedded. Weathers lead-gray	19	5.8
Dolomite and limestone: white-gray and gray, coarsely crystalline, thick-bedded dolomite. Weathers buff-tan as underlying, and intercalated intervals of dove-gray, fine-grained limestone. Much black and white mottling in many dolomite beds	49	15.1
Dolomite and limestone: white, and dull- and buff-gray, extremely coarsely crystalline thick-bedded dolomite occasionally mottled dark and light gray. A thickness of 30 to 40 feet of dull-gray thinner-bedded fine-grained limestone in middle part of this interval. Some beds of dove-gray, fine-grained (nearly lithographic), thin-bedded limestone in upper fourth	43	13.2
Interval covered with talus or float of dolomite dull-gray in lower three-fourths, pale-buff in upper fourth, all coarsely to medium crystalline. Forms buff-tan broken cliffs elsewhere	74	22.8
Limestone: black-gray, medium to coarsely crystalline, dolomitic, in 1- to 7-inch beds. Weathers dark-gray. This limestone appears to be an "island" surrounded by dolomite	9	2.8
Dolomite: white-gray, coarsely crystalline, massive, thick-bedded (1½ to 3 feet). Weathers light-buff-gray and forms lowest prominent cliffs in Cathedral formation	72	22.1
Dolomite: white and dark-gray mottled, coarsely crystalline. Weathers pale-tan-gray and forms low cliffs	10	3.1
Dolomite: white- and pale-buff-gray, medium crystalline (sugary) thin-bedded (1 to 4 inches). Weathers pale buff-gray and to sugary surfaces. Contains 1/20- to 1/5-inch veinlets and stringers of white crystalline dolomite in lower beds and forms abrupt contrast with underlying lead-gray limestone	24	6.8
Total thickness of Cathedral dolomite	1085	335.8

[3] In his earlier publications the writer accepted Dr. C. E. Resser's incorrect spelling *"Ptarmingia"* for *Ptarmigania*.

Ptarmigan? limestone	Feet	Meters
Limestone: black-gray, fine-grained, more or less vitreous, in 1- to 2-inch beds. Contains flakes and blebs of limonitic clay. Weathers dull-gray and forms top of lead-gray cliffs and of Ptarmigan? limestone	27	8.3
Limestone: black- and pale-sooty-gray, fine-grained, argillaceous, hard, thin-bedded (1 to 5 inches, average 2½ inches). Contains blebs and flakes of limonitic orange-tan and some salmon-colored clay. Weathers dark- or lead-gray stained orange and pale-red with clay, and to rough irregular plates and angular fragments. Looks like typical Ptarmigan limestone in type section	33	10.1
Limestone: dull-gray, slightly oölitic and strongly banded with tan-weathering clay in lower 6 feet and black-gray, extremely fine-grained, thin- and irregular-bedded hard limestone in 1- to 2½-inch beds. Contains many flakes and blebs of tan and salmon-red clay parallel to bedding	35	10.8
Limestone: white-gray, fine-grained, thin-bedded, finely arenaceous (calcarenite), and partings of calcareous clay. Weathers drab-buff similar to top of Mount Whyte formation on Ptarmigan Peak	5.5	1.7
Limestone: light-gray, fine-grained, argillaceous, massive-bedded. Contains much clay in form of blebs and nodules. Weathers drab-tan. Overlain by dull- and dark-gray finely oölitic, thick-bedded limestone which weathers rusty tan and forms massive cliffs and breaks down to angular fragments. Similar to upper part of Mount Whyte formation on Ptarmigan Peak	14	4.3
Limestone: pale-blue-gray, fine-grained. Contains specks of limonite, trilobite fragments, and numerous blebs and flakes of gray clay. Weathers tan-gray	7.5	2.3
Limestone: dull- and dark-gray, finely to coarsely oölitic in some beds, others finely crystalline; massive, hard, in 2- to 10-inch beds. Stylolites well developed in basal massive part of interval. Weathers drab-gray and forms cliffs. Some beds contain fragmentary fossils	27	8.3
Interval covered, underlain by oölitic and shaly limestone	22	6.8
Limestone: dark- and little tan-gray, finely to medium oölitic, in 1- to 6-inch beds, interbedded with pale-gray fine-grained limestone which contains flakes and nodules of pale-buff clay. Darker gray beds contain trilobites and brachiopods, and an occasional bed contains large pisolites or *Girvanella* as much as ⅜ inch in diameter. Oölitic beds weather dark gray. Fossil loc. C6d in upper third of interval	22	6.8
Fossils: *"Alokistocare"* 2 n. sp. *"Antagmus" skapta* (Walcott) *Kochaspis carina* (Walcott) *K.* cf. *cecinna* (Walcott) n. gen. of trilobites *Scenella* n. sp. cf. *S. reticulata* Billings *Stenothecoides?* sp. *Hyolithes* sp.		
Limestone: one bed: black-gray, finely to coarsely oölitic upward (½ to 5 mm. in diameter). Weathers black-gray and forms conspicuous marker in cliffs	2	0.8
Total thickness of Ptarmigan? limestone	195	60.0
Total thickness of Middle Cambrian	2860	882.5

LOWER CAMBRIAN	Feet	Meters
Mount Whyte formation		
Limestone: dark-gray, medium crystalline to fine-grained, in ½- to 7-inch beds, finely to coarsely oölitic. Contains trilobite fragments. Interbedded with tan-steel-gray, finely crystalline, slightly vitreous, platy limestone in ¼- to 1½-inch beds. Contains limonite specks and some metamorphosed green shale partings. Interval weathers dark gray	12	3.7
Limestone: pale- and black-gray, fine- to medium-grained, massive, in ½- to 3-inch beds which weather to platy slabs. Contains a few thin (1- to 2-inch) beds of poorly developed oölite. Weathers drab-gray and forms a bench of broken ledges. In upper 10 feet fossil loc. C6c. *Plagiura* fauna	22	6.8
Fossils: *Inglefieldia* n. sp.; *Kochiella crito* (Walcott); *Onchocephalus?* sp.; *Plagiura cercops?* (Walcott); *Plagiurella cleadas* (Walcott); *Poulsenia?* n. sp.; *Prozancanthoides* sp.; *Ptychoparella* 3 n. sp.; *Zacanthopsis?* sp.; *Hyolithes* sp.		
Limestone: dull- and pale-blue-gray grading up into pale-salmon-gray, fine- to medium-grained, argillaceous, irregular-thin-bedded, massive. Contains flakes and nodules of calcareous green and gray clay and irregular bands of arenaceous clay in upper part which weathers rusty and to raised bands. Forms sheer cliffs which weather to rounded "pillowy" ledges as on Ptarmigan Peak	64	19.7
Limestone: as in underlying interval, but less slaty cleavage developed and less argillaceous. Weathers buff-tan stained pale-red	19	5.8
Limestone: pale-gray, argillaceous, slaty, as in underlying interval. Weathers to large cavities in cliff faces	22	6.8
Limestone: dull-pale-gray, fine-grained, argillaceous, thin- and irregular- (flaggy) bedded, strongly cleaved at 34 degrees to bedding planes by slaty cleavage which causes limestone to appear as shale from a distance or on superficial observation. Weathers pale-green-gray and to rusty brown on some cliff faces. Strike and dip of cleavage N. 39° W., 45° W. Fossil loc. C6b 16 feet above base of interval	47	14.5
Fossils: *Inglefieldia?* sp. indet.; *Kochiella* sp.; *Plagiura?* sp.; *Proliostracus* n. sp.; *Prozancanthoides* sp.; *Ptychoparella* 2 or 3 n. sp.; n. gen. of trilobites; brachiopods, several species; *Hyolithes* sp.		
Shale and limestone: green, argillitic, chunky, very slightly calcareous shale in 6- to 24-inch intervals, interbedded with dull-gray, argillaceous fine- to medium-grained thin-bedded massive limestone which contains trilobite hash beds. Here fossil loc. C6a. Rocks weather dull-green stained tan and form a cliff above the purple-maroon cliffs at the top of the St. Piran	49	15.1
Fossils: *Bonnia columbensis* Resser; *B. fieldensis* (Walcott); *Olenellus* sp.; *Onchocephalus?* sp. indet.; *Nisusia?* sp.		
Total thickness of Mount Whyte formation	235	72.4

St. Piran sandstone	Feet	Meters
Sandstone: purple-maroon, coarse- and fine-grained, thick- and irregular-bedded, strongly cross-bedded, all more or less calcareous. Weathers purple-maroon and forms sheer cliffs at top of St. Piran sandstone	12	3.7
Note: The contact between the Mount Whyte and St. Piran formations is not clearly visible from a distance.		
Shale and sandstone: pale-green, argillitic, slaty shale in 1- to 3-foot intervals, interbedded with brown, calcareous, coarse-grained, massive, cross-bedded sandstone which contains much limonite and tiny quartz pebbles and angular grains and some hematite. Sandstones irregular-bedded and lenticular, and occasionally ripple marked	18	5.5
Shale, sandstone, and quartzite: pale-green, finely arenaceous, argillitic, chunky shale interbedded with purple-maroon hematitic and red-tan coarser-grained sandstone. Forms bench between cliffs on southeast side of Mount Bosworth	48	14.7
Quartzite and sandstone: purple and maroon, medium- to fine-grained, vitreous, hard, thick- and thin-bedded quartzite, platy-bedded and more or less argillitic in upper part of interval and interbedded with much coarse-grained tan-gray, cross-bedded sandstone in lower half. Much hematite, limonite and tiny quartz grains irregularly disseminated. Weathers purple-maroon and forms sheer cliffs	54	16.6
Sandstone: dark-red, coarse- to medium-grained, hematitic, thick-bedded more or less soft. Weathers dark maroon and forms broken ledges	34	10.5
Shale and sandstone: pale-green, slaty, more or less fissile shale in 12- to 15-inch intervals, interbedded with two 14-inch beds of calcareous sandstone which contain finely disseminated limonite. Shale weathers green; sandstone weathers drab-tan	8	2.5
Quartzite and sandstone: maroon-purple and dark-red, coarse-grained, thick-bedded sandy quartzite and sandstone. Contains much hematite in flakes and grains. Weathers deep maroon and forms lower part of dark-red band in cliffs	58	17.8
Quartzite: purple and maroon, and some steel-blue-gray, medium-grained, thick-bedded. Some beds strongly banded purple and brown. Weathers dark-purple and forms ledges. Interval much covered locally, but forms part of sheer cliffs to east	43	13.2
Quartzite: purple-maroon, thick-bedded, medium-grained, vitreous. Weathers purple and forms ledges	30	9.2
Quartzite: pink-tan and pale-maroon-buff, fine- to medium-grained, hard, thick-bedded, some beds banded purple. Weathers pale-red and forms sheer cliff here	32	9.8
Quartzite: maroon and red, medium- and coarse-grained, vitreous, hard, in 1- to 5-foot beds. Weathers maroon stained pale-red. Upper part forms bench beneath pink sheer cliffs here	28	8.6
Quartzite: one bed: dull-maroon-purple, medium-fine-grained, hard, vitreous. Weathers dark-purple and to smooth rounded joint faces, and forms conspicuous persistent marker in section	9	2.8
Quartzite and sandstone: similar to underlying quartzite, but less massive, softer, thinner-bedded, and coarser-grained. Forms bench beneath dull-purple-maroon ledge	15	4.6
Quartzite: tan and tan-gray, hard, vitreous, massive, in ½- to 6-foot beds, strongly cross-bedded locally. Contains occasional lenticular bed of coarse-grained, pale-gray sandstone and one or two 18-inch intervals of dark-green, chunky, argillitic shale. Weathers pale-tan stained pink and tan in irregular spots, and forms conspicuous low cliff-ledges in section	94	28.9
Quartzite: one bed: tan and brown-gray, medium-grained vitreous quartzite. Weathers tan stained brown. Joint face at top of bed contains ⅛- to ¼-inch quartz crystals	6	1.8

	Feet	Meters
Shale and sandstone: green slaty shale and coarse-grained, thin-bedded sandstone. Much covered and forms slope here	18	5.5
Quartzite and shale: gray and tan, thick- and thin-bedded quartzite, and occasional ½- to 2-foot interval of dull-green shale	40	12.3
Quartzite and sandstone: tan, gray, and red, coarse-grained, vitreous, thick-bedded quartzite and quartzitic sandstone. Interval partly covered	65	20.0
Lake Louise? shale		
Sandstone and shale: gray and tan quartzitic sandstone in 4- to 12-inch beds, and intercalated green slaty shale as below in section	20	6.2
Quartzite and shale: tan-gray, massive, cross-bedded, more or less vitreous quartzite in 1- to 2½-foot beds. Weathers tan on cliff faces. Several 1- to 6-inch intervals of green shale intercalated in upper part of interval	24	7.4
Shale and quartzite: dull-green, slaty shale in 1- to 8-foot intervals, interbedded with 1- to 2-foot beds of coarse-grained, gray, tan, and some purple quartzite. May be upper part of Lake Louise shale member of Fort Mountain formation	27	8.3
Quartzite and shale: quartzite thick- and irregular-bedded as below in section, and a few thin intercalated green slaty shale partings	12	3.7
Shale and quartzite: green, fine-grained, slaty shale in 1- to 5-foot intervals, interbedded with 6- to 24-inch beds of gray and tan, coarse-grained, arenaceous quartzite spotted red on some beds, and occasionally cross-bedded. Shale weathers dull-tan-green; quartzite weathers red-tan and gray	41	12.6
Shale: dull-green, metamorphosed, slaty, massive, weathers dark-green stained brown and tan by limonite. Cleavage poorly developed. May form base of Lake Louise shale	6	1.8
Total thickness of Lake Louise? shale	130	40.0
Quartzite, sandstone, and shale: gray, medium-coarse-grained, cross-bedded, arenaceous quartzite. Weathers buff-gray. Gray coarse-grained sandstone in extremely irregular beds 1 to 12 inches thick which are lenticular and ripple-marked. All interbedded with dull-olive-green slaty, fissile shale in 1- to 4-inch beds. Slaty cleavage 15 to 20 degrees to bedding surfaces of competent beds	8	2.5
Total thickness of St. Piran sandstone	750	230.5
Total thickness measured of Lower Cambrian	985	302.9
Total thickness measured of Mount Bosworth section	5105	1572.9

COMPARISON OF ORIGINAL, ALTERED, AND EMENDED MOUNT BOSWORTH SECTIONS

The Upper Cambrian formations above the Arctomys were not measured. Consequently the upper part of the original Mount Bosworth section cannot be discussed here other than to note that with the strike and dip of the beds on Paget Peak only 900 to 1200 feet of thickness is possible between the Bosworth-Arctomys contact and the top of Paget Peak (Pl. 2, fig. 1) which Walcott (1908b, p. 204) stated was formed of the highest beds in his section. The combined Bosworth, Paget, and Sherbrooke formations, therefore, are probably more nearly 1200 than 2307 feet in thickness.

The Arctomys formation on the ridge connecting Mount Bosworth and Paget Peak is extremely difficult to measure accurately because on the crest of the ridge the beds were displaced by ice, the southwest side of

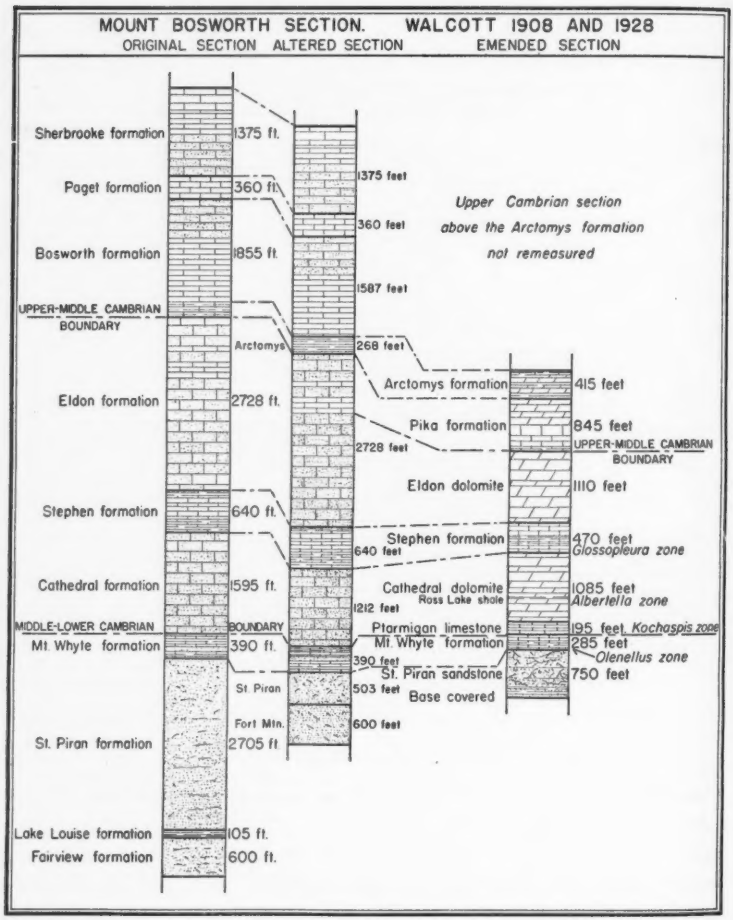

FIGURE 2.—*Comparison of original altered, and emended Mount Bosworth sections*

the ridge is a series of broken dip-slopes which are much covered with talus, and, although the east and northeast sides of the ridge are cut into sheer cirque walls, at no one place is the entire formation available.

The most reliable figure obtained in the field (415 feet) is a combination of direct measurement, computation from instrument readings, and, in the upper 195 feet, a number of estimates. In spite of the doubt as to the accuracy of the thickness given in the emended section (Fig. 2) the Arctomys is considerably thicker than 268 feet (Pl. 2, fig. 1).

One of the most obvious discrepancies in the altered section (Walcott, 1928, p. 311, Pl. 67) was the statement that the Arctomys formation rests upon the Eldon, whereas in the photograph of the south side of Mount Bosworth the beds resting on the Eldon and forming the top of the mountain were labeled "Bosworth." The beds which rest on the Eldon as illustrated on Plate 67 of that paper belong neither to the Arctomys nor Bosworth but were included in the upper 963 feet of the Eldon formation in Walcott's original section (Walcott, 1908b, p. 208-209; 1928, p. 311). These beds actually total 835 feet in thickness and are typical Pika dolomites, limestones, and shales (Deiss, 1939a, p. 1008-1009) instead of any one of the three formations to which they had been previously assigned.

Another difference between the original and emended sections is the presence in the latter of the Pika formation (Fig. 2). Although the rocks were thoroughly searched, fossils could not be found in the formation. However, these rocks are similar lithologically to those in the type section of the Pika on Castle Mountain and are sharply demarcated by color and composition on Mount Bosworth from the underlying Eldon dolomite (Pl. 1) and overlying Arctomys shales (Pl. 2, fig. 1). Therefore, they form a readily mappable unit in this area and occupy the same stratigraphic position as on Castle Mountain and Ptarmigan Peak.

Fossils could not be found above or below the Pika in the Arctomys and Eldon formations. Consequently, faunal evidence was not available for locating the Upper-Middle Cambrian boundary on Mount Bosworth. Because the 835 feet of strata below the Arctomys is lithologically equivalent to the Pika which probably is the basal Upper Cambrian formation in this part of the Canadian Rockies (Deiss, 1939a, p. 1015-1016), the Upper-Middle Cambrian boundary is tentatively placed between the Pika and Eldon formations on Mount Bosworth (Pl. 1). As this 835-foot interval was included in the upper 963 feet of the Eldon by Walcott, the Upper-Middle Cambrian boundary is drawn 835 feet lower in the emended section than in the original.

When one realizes that the maximum relief along the strike of the beds between the floor of Kicking Horse Valley and the top of Mount Bosworth is only 4000 feet, Walcott's photograph of the south side of the mountain (Walcott, 1917b, Pl. 1; 1928, Pl. 67) shows at a glance that the Eldon dolomite forms less than a third of the thickness of rocks

which comprise the mountain. For several years the thickness—2728 feet—has been known, therefore, to be erroneous. Even if the 963 feet of Walcott's Eldon (now included in the Pika) be added to the 1110 feet of beds labeled "Eldon" on the photograph, the formation would be only 2073 feet instead of 2728 feet thick, of which possibly 1400 feet is shown on the photograph (Walcott, 1928, Pl. 67). This is the largest error in measurement among the Lower and Middle Cambrian formations in the section.

The Stephen formation also is much thinner than originally was stated, being 470 feet instead of 640 feet thick. Because the Stephen changes laterally from shale to impure limestones and is transitional into the overlying Eldon, especially in the western part of Mount Bosworth where the formation is also largely concealed by forests and rock mantle, the discrepancy in thickness between the original and emended section is understandable if the formation was measured in this area on the lower slopes of the mountain. The upper and lower boundaries of the Stephen formation, however, are clearly defined and well exposed above 7300 feet in elevation on the side of Mount Bosworth at a point S. 36° W. of Sink Lake.

Probably the most inexplicable error in the original section was the failure to find the *Albertella* zone exposed in the Cathedral formation (Pl. 3, fig. 1), with the result that this trilobite was assigned to the Lower Cambrian Mount Whyte formation (Walcott, 1908b, p. 213-214). This error becomes more confusing because the precise stratigraphic position of the *Albertella* zone in the Cathedral formation on Castle Mountain was known in 1907 (Walcott, 1917b, p. 12), the same year that the Mount Bosworth section was first measured. In 1915 Burling located the Ross Lake (*Albertella*) shale in the Cathedral "limestone" on Mount Bosworth, described its position and the geographic location of the outcrop, and the following year published a photograph of it (Burling, 1916, p. 469-470, Fig. 1). The same summer Walcott found the fauna in place in the Ross Lake section and reported it to be in place on the south side near the base of Mount Bosworth (Walcott, 1917b, p. 12, Pl. 1). In spite of all the field information known by 1928, the *Albertella* zone was again indicated incorrectly in the Ptarmigan beneath the Cathedral formation on Mount Bosworth (Walcott, 1928, Pl. 67 [4]), and the stratigraphic position of *Albertella* was not even mentioned in the section except in a footnote. The persistent failure to correct the position of the Ross Lake shale in the Mount Bosworth section in incomprehensible. The shale is sharply demarcated from and clearly

[4] This photograph is the one which Walcott first published in 1917 and on it the same incorrect position of the *Albertella* shale is given on the east ridge of Mount Bosworth.

FIGURE 1. CONTACT OF BOSWORTH ON ARCTOMYS, AND ARCTOMYS ON PIKA
Ridge between Mt. Bosworth and Paget Peak.
B = Bosworth, A = Arctomys, P = Pika formations.

FIGURE 2. DIFFERENTIAL WEATHERING OF ARENACEOUS INCLUSIONS IN UPPER PIKA DOLOMITE
On ridge between Mt. Bosworth and Paget Peak.

FIGURE 2. TYPE SECTION OF NAISET FORMATION
On east side of Naiset Point. C = Cathedral limestone, N = Naiset formation, G = Gog sandstone, W = Wonder Pass.

FIGURE 1. PART OF OUTCROP OF ROSS LAKE SHALE
On east ridge of Mt. Bosworth. Length of hammer handle 15 inches. C = Cathedral dolomite.

ROSS LAKE SHALE AND NAISET POINT

visible in the Cathedral dolomite (Pl. 3, fig. 1) and is almost completely exposed for nearly a quarter of a mile on the crest of the east spur of Mount Bosworth where the field relationships are simple and readily obtained. The Ross Lake shale is given in the emended section at least 385 feet higher stratigraphically than in Walcott's section of 1928 and approximately at the horizon in the Cathedral dolomite reported by Burling (1916, p. 469-470).

The 195 feet of dark-gray thin-bedded limestone doubtfully assigned to the Ptarmigan formation in the emended section is clearly the upper 120-foot interval of Walcott's Mount Whyte formation (Walcott, 1928, p. 313) but has no equivalent in the lower part of his Cathedral formation. These beds are also the ones marked "Ptarmigan" on his photograph (Walcott, 1928, Pl. 67). Consequently, if these dark-gray-weathering limestones are the equivalent of the Ptarmigan formation they clearly are not included in Walcott's altered Cathedral formation on Mount Bosworth as stated by Resser (Walcott, 1928, p. 312-313).

The Lower Cambrian quartzites, sandstones, and interbedded shales which constitute the lower part of the Mount Bosworth section cannot be subdivided into formations, and probably all are equivalent to the upper part or most of the St. Piran sandstone. The beds doubtfully labeled Lake Louise shale in the emended section may be only a local argillaceous phase of the St. Piran, and their color, manner of weathering, and the topography produced are not from any distance sufficient to demarcate them from the overlying 610 feet of beds. The St. Piran on Ptarmigan Peak, Castle Mountain (Deiss, 1939a, p. 971-972, 982-983), and Mount St. Piran (Walcott, 1908b, p. 216) is thicker than the entire exposure on the south side of Mount Bosworth. These facts support the conclusions that all the clastic rocks exposed beneath the Mount Whyte formation in the Mount Bosworth section are equivalent to the St. Piran sandstone to the south and southeast and that the Lake Louise shale member and the Fort Mountain sandstone are not exposed on the south side of Mount Bosworth.

The total thickness of rocks measured on Mount Bosworth in 1939 is 5105 feet, to which may be added the rough estimate of 1200 feet for the combined Sherbrooke and Paget formations. Thus the section totals approximately 6305 feet instead of 12,353 feet in thickness (Walcott, 1908b, p. 217). Likewise the thickness of the Lower and Middle Cambrian formations on Mount Bosworth are normal for the area, being essentially equivalent to those in the other sections measured to the south and east. The great discrepancies in thickness of such formations as the Cathedral and Eldon reported from different sections therefore do not exist and need no special explanations.

MOUNT ASSINIBOINE AREA, BRITISH COLUMBIA

GENERAL STATEMENT

The Mount Assiniboine section probably was first measured in 1916 (Walcott, 1917b, p. 12) but was not published until 1928 (Walcott, 1928, p. 296-299). The writer's field work in Banff National Park in 1938 (Deiss, 1939a, p. 954-955) showed that the Mount Whyte and Ptarmigan formations thin southward from Ptarmigan Peak to Castle Mountain; that the *Albertella* zone is above the base of the Cathedral dolomite instead of in the Ptarmigan limestone as had been stated (Walcott, 1917a, p. 14-15); and that the Middle-Lower Cambrian boundary is marked by *Kochaspis cecinna* (Walcott) and *K. gogensis* (Walcott), forms which were placed in the Lower Cambrian Mount Whyte formation (Walcott, 1917b, p. 65).

Lower Cambrian sediments are absent in northwestern Montana, and Lower Cambrian fossils are unknown from Montana, Idaho, and extreme northern Utah. Consequently the Waucobian seas must have invaded the Cordilleran trough from the north and south but did not become confluent during the epoch. If this is true, the Lower Cambrian formations should have thinned southeastward from Castle Mountain toward the Assiniboine region. Walcott (1915, p. 197-198) was the first to note this land barrier during the Cambrian in the Cordilleran trough and gave to it the name Montana Island. The conditions on the south side of the "island" during Middle and Upper Cambrian time were recently described in connection with the Cambrian paleogeography of Montana (Deiss, 1939b, p. 58-61). In light of the information available by 1939 the Assiniboine section seemed likely to give additional evidence on the position of the south shore of the Lower Cambrian sea in Canada and also on the character of the northern part of the Montana Island.

The Mount Whyte formation in the original Assiniboine section was said to be 785 feet thick (Walcott, 1928, p. 297-298). This thickness is abnormal since the Mount Whyte thins southward from Ptarmigan Peak to Castle Mountain (Deiss, 1939a, p. 1017, Fig. 5). On the other hand *Kochaspis cecinna* (Walcott) was listed in the upper 23 feet of the Mount Whyte in the Assiniboine section, and *K. gogensis* (Walcott) was described from the same locality (62w) in 1917 (Walcott, 1917c, p. 88). The only other trilobite given from the Mount Whyte in the original Assiniboine section was *"Alokistocare" cleora* (Walcott). These three trilobites are Middle Cambrian in age, and *Kochaspis cecinna* and *K. gogensis* mark the basal beds of that series (Deiss, 1939a, p. 980, 1012-1015). The only Lower Cambrian trilobite reported from the Assiniboine section was *Olenellus* sp. of which several fragmentary cranidia were observed by Walcott in his St. Piran formation (upper Gog) at least 235

feet below its top.[5] However, the *Olenellus* zone in the Castle Mountain, Ptarmigan Peak, Ross Lake, and Mount Bosworth sections is in the basal limestones of the Mount Whyte formation. The rocks constituting the Mount Whyte formation in the original Assiniboine section, therefore, seemed to be Middle Cambrian and probably temporally equivalent to the Ptarmigan limestone on Ptarmigan Peak, whereas the St. Piran formation appeared to be of the same age as the Mount Whyte in its type locality.

In order to obtain additional evidence concerning these problems, Walcott's Mount Assiniboine section was remeasured in 1939. The rocks were carefully searched for fossils which were accurately zoned in the newly measured section, and the revised formations were mapped with a plane table to test their usability as cartographic units.

Mount Assiniboine forms the center of an open, gently plunging syncline, contains the youngest strata in the structure, and is the highest point in elevation on the divide between Alberta and British Columbia in the southern Canadian Rockies. The peak is buttressed on the northeast by a great ridge consisting of Mount Magog and Mount Terrapin, and on the northwest by another ridge on which Mount Sturdee and Wedgwood Peak are the highest points. The eastern ridge bifurcates north of Mount Terrapin. One part extends east to Mount Towers and thence northeast to Wonder Pass; the other continues nearly due north and terminates in the sheer-walled narrow spur which Walcott (1928, p. 296) named Naiset Mountain.[6] Mountain Assiniboine, together with these ridges and the one to the south which ends in Mount Aye, forms a huge, high, more or less isolated mountain mass which Walcott (1928, Pl. 61) called the Assiniboine Massif.

Structurally and stratigraphically the Assiniboine "Massif" is sharply separated from the rocks in the ranges to the north and east. The basal Cambrian conglomerate rests disconformably upon the black, fissile, tan-weathering Hector shale which crops out in a narrow belt of varying width on the north and east sides of the mass where the shale and overlying Cambrian rocks have ridden over Devonian (?) limestones and shales along a low-angle thrust. The larger and more apparent structural relationships in this part of the Canadian Rockies are similar to those in the province east of Bow Valley where the Hector shale and superjacent Cambrian sediments also form a gentle syncline and were thrust eastward over steeply dipping and slice-faulted (high-angle) younger Paleozoic limestones (Deiss, 1939a, p. 962). These structural relationships are

[5] Apparently the specimens were too poorly preserved to be collected. At least the collections of the United States National Museum contain no example of *Olenellus* from the Mount Assiniboine area.

[6] Naiset Mountain is labeled Naiset Point (Pl. 5, fig. 1) on the official Canadian topographic maps.

astonishingly similar to those in northwestern Montana south of Glacier Park where the gently dipping Beltian and Paleozoic rocks in the Lewis and Clark Range have been thrust eastward upon the slice-faulted Jurassic and Lower Cretaceous shales and sandstones which form the Sawtooth Range (Clapp, 1932, p. 25; geologic map).

LOCATION OF AREA

The Assiniboine "Massif" forms more than 4 miles of the Continental Divide in the southeast part of the region (Fig. 1). The principal part of the mass is northwest of the Divide in the provincial Mount Assiniboine Park of British Columbia. The parallel 50° 52′ N. Lat. and the meridian 115° 39′ W. Long. intersect near the apex of Mount Assiniboine which is 21 miles S. 12° W. of the Government Building in Banff, Alberta, and approximately 130 miles due north of the International Boundary. Mount Assiniboine is covered by sheet number 12 of the Interprovincial Boundary Commission, a three-color topographic map on the scale of 1/62,500 and the contour interval of 100 feet.

MOUNT ASSINIBOINE SECTION

Neither the original nor the emended Mount Assiniboine sections was measured on Mount Assiniboine, and only the estimated upper part of the latter section refers to beds on that peak. The precise location of the traverse along which the original section was measured is unknown. Concerning the location of his section Walcott (1928, p. 296) wrote:

> "The measured section is from the summit of the ridge down to the level of Gog Lake (7,200 feet, 2,194.6 m., above sea level). The summit of the ridge is formed of massive-bedded arenaceous limestones of the Cathedral and Ptarmigan formations. About 400 feet (121.9 m.) of the arenaceous limestone remains on the north section of the ridge, and 1500 feet (457.2 m.) at Naiset Mountain."

On the same page Walcott gave a graphic profile section of the ridge between the top of Mount Assiniboine and the north end of Wedgwood Peak. This ridge is nearly 2 miles west of Naiset Point and Gog Lake. Because Walcott's locality (62w) is described as ". . . above Gog Lake, below Wonder Pass, on Continental Divide, in British Columbia, 19 miles (30.4 km.) southwest of Banff, Alberta, Canada," part of the original section also must have been obtained on the northeast spur of Mount Towers (Walcott, 1917b, p. 68).

The emended section (Fig. 3) was measured from a point in the small steep-walled canyon on the southeast side near the top of Wonder Pass at an elevation of approximately 7800 feet where the basal Gog conglomerate rests on the pre-Cambrian Hector shale (Pl. 4, fig. 2). The section traverse crosses Wonder Pass and continues up the Continental Divide on the northeast spur of Mount Towers to the base of the Naiset formation. On this ridge the Naiset strata are much sheared and dis-

torted and are partly covered with talus from the overlying Cathedral dolomite (Pl. 2, fig. 1) to the extent that a dependable section cannot be obtained there. The section was carried west to Naiset Point where it was "tied in" on the contact of the platy argillaceous brown and gray basal Naiset sandstones against the pale-tan massive upper Gog quartzite. The upper part of the emended section was then measured on the east side of Naiset Point (Pl. 3, fig. 2) where the rocks comprise the Naiset formation and the lower part of the Cathedral dolomite. The emended section lies in the N. ½ sec. 21 NW. ¼ sec. 22, and the SW. ¼ sec. 28, T. 22 N., R. 12 W. The part of the traverse in sec. 22 is in Alberta, and the part in secs. 21 and 28 is in British Columbia.

When the writer was in the area the rocks above the basal 257 feet of the Cathedral dolomite could not be climbed because of snow slides and falling ice. The rocks of the Stephen, Eldon?, and Upper? Cambrian formations were not measured. The estimated thicknesses of these formations given in the emended section were obtained by observation from different points in the field, by computing the thicknesses from the strike and dip of the surrounding region, and by scaling the width of outcrops on the areal geologic map. Because well-preserved specimens of *Glossopleura* are common in the moraines at the head of Magog Lake, the Stephen formation is certainly present. The identification for areal mapping of the Eldon? and unnamed Upper? Cambrian formations was based on color, characteristic weathering, and the type of topography which they produce. Obviously these formations are of little or no value for careful stratigraphy. They are given solely because they indicate for future workers the probable age of the rocks in the upper part of the Assiniboine section and show how small a part has been accurately measured.

	Feet	Meters
Upper? Cambrian limestones and dolomite		
Limestone and dolomite: thick- and thin-bedded limestone forming sheer cliffs and benches on upper part of Mount Assiniboine. Section only observed from glacier at south end of Magog Lake and from top of Naiset Point (estimated)	1300	400.0
Total estimated thickness of Upper? Cambrian	1300	400.0
MIDDLE CAMBRIAN		
Eldon? formation		
Dolomite and limestone: white-gray and dull-gray, fine-grained and crystalline, thick-bedded dolomite and some thin-bedded tan-weathering limestone forming benches between cliffs on lower part of Mount Assiniboine proper (estimated)	1200	369.0
Total estimated thickness of Eldon? formation	1200	369.0

Stephen? formation	Feet	Meters
Limestone and shale: dark-gray, fine-grained, platy-bedded, fossiliferous limestone and interbedded intervals of green-gray, fine-grained, argillaceous, shaly-bedded, massive limestone and shale which contains *Elrathina* cf. *cordillerae* (Rominger). In lower part thin-bedded dull-gray limestone which contains *Glossopleura* sp. Weathers dark-gray and buff and forms top of Terrapin Peak. Description and fossils taken from rocks in moraine south of Magog Lake (estimated)	250	76.9
FOSSILS: *Elrathina* cf. *cordillerae* (Rominger) *Glossopleura* sp. *Solenopleurella* sp. *Acrotreta* sp.		
Total estimated thickness of Stephen? formation	250	76.9
Cathedral formation		
Dolomite and limestone: tan, gray, and black, massive-bedded, coarse- to fine-grained, cliff-forming dolomite interbedded with irregular intervals of dark-gray thinner-bedded limestone. In upper-middle part one interval 50 to 75 feet thick of dull red-weathering dolomite which is black-gray on fresh fracture. Estimated by scaling planetable map	743	228.2
Note: Here top of Walcott's Ptarmigan formation (Fig. 3).		
Limestone and dolomite: black-gray, fine-grained, fossiliferous, thin- and thick-bedded limestone. Sooty-black, slightly argillaceous, thin-bedded (¼ to 4 inches, average 1 inch), fine-grained limestone in upper 25 to 40 feet. Weathers dull-gray and contains *Albertella*. Top of fossil loc. C9c. Locally limestone grades laterally into buff-gray and gray massive-bedded limestone which contains flakes and nodules of drab-gray clay, and elsewhere into tan and gray sugary massive-bedded dolomite	50	15.4
Limestone and dolomite: black-gray, extremely fine-grained to medium crystalline, platy-bedded, fossiliferous limestone which weathers black-gray. Contains irregular reeflike masses of gray limestone which contains white calcite stringers and veinlets which weather buff. Local masses of mottled buff and gray, sugary, thick-bedded dolomite which weathers buff-tan. Base of fossil loc. C9c in platy limestones	59	18.1
FOSSILS: *Albertella nitida* Resser *A.* n. sp. cf. *A. helena* Walcott *A.* sp. *"Antagmus" skapta* (Walcott) *"A"?* 2 n. sp. *Periomma?* 2 n. sp. *Poliella* cf. *P. prima* Walcott *P.* n. sp. *Ptarmigania* cf. *P. rossensis* (Walcott) *P.* sp. *Zancanthoides* n. sp. *Hyolithes* sp. Contains small cranidia of n. gen. known from Comet shale of Nevada.		
Limestone and dolomite: sooty-gray, finely to medium crystalline, thin-bedded (1 to 8 inches). Some beds contain fossils, others smaller flakes and nodules of buff clay. Weathers dark-gray and forms sheer cliffs. Fossil loc. C9c—loose. Cathedral limestone on Naiset Point	70	21.5
FOSSILS: n. gen. cf. *Kochaspis* Resser *Kochina?* n. sp. n. gen. cf. *Onchocephalus* Resser *Poliella* 2 n. sp. *Ptarmigania* sp. *P.?* n. sp.		

	Feet	Meters
Limestone: dull- and black-gray, medium to coarsely crystalline, fossiliferous, strongly oölitic in many beds, thick-bedded (3 to 60 inches). Weathers dull-gray and to irregular beds. Contains *Kootenia* and other trilobites in lower third. Fossil loc. C9b. In upper two-thirds are 8- to 18-foot intervals of buff and gray salt and pepper and banded, medium crystalline sugary thick-bedded dolomite which grades laterally into dark-gray pure fossiliferous limestone	56	17.2
Fossils: *Kootenia* cf. *exilaxata* Deiss *K.* n. sp. *"Onchocephalus" thia* (Walcott) *"O."* n. sp. n. gen. cf. *Syspacephalus charops* (Walcott) *Helcionella?* sp. *Micromitra?* sp. *Scenella?* n. sp. similar to sp. in loc C6d *Stenothecoides?* sp. *Hyolithes* sp.		
Limestone: dull-gray, fine- to medium-grained, platy, grading up into massive, thick beds whose hemispheric upper surfaces resemble algal reefs. Contains bands and irregular masses of bright-orange-tan limonitic clay. Weathers pale-gray stained buff-tan	3.5–4	1.1–1.2
Limestone: dark-blue-gray, fine- to medium-grained, thick- and thin-bedded, massive. Contains large (½ to 1 inch) *Girvanella*, much limonite, and fragments of trilobites, and flakes and nodules of bright orange-tan-weathering clay. Limestone weathers gray	18	5.5
Total thickness of Cathedral formation	1000	307.1

Naiset formation (type locality)

Shale, conglomerate and limestone: bright-tan, finely arenaceous, calcareous, platy-bedded (1/64 to 1/8 inch) to fissile shale. Weathers bright-tan, and intercalated 2- to 11-inch beds of orange-tan intraformational conglomerate which contains tiny (⅜ to 1 inch, average ¾ inch) green limestone pebbles and occasional cranidia and pygidia. Much tan limonite. Upper two-thirds of interval dull-gray, fine-grained, argillaceous limestone streaked with tan limonite bands 1/64 to 1/8 inch wide. Weathers drab-tan-gray and to smooth beds and contains occasional cranidia. Top of fossil loc. C9a. Many elongate-oval *Girvanella?* in conglomerate beds	22	6.8
Fossils: *"Amecephalus" cleora* (Walcott) *"A."* 3 n. sp. n. gen. cf. *Amecephalus* Walcott n. gen. cf. *Elrathia* Walcott *Kochaspis* n. sp. cf. *K. chares* (Walcott) *K.?* n. sp. cf. *K.? gogensis* (Walcott) n. gen. cf. *Kochiella* Poulsen *Poliella?* sp. indet. *P.?* n. sp. n. gen. of trilobites *Nisusia?* 2 sp. *Wimanella?* sp. indet. *Hyolithes* sp.		
Limestone and mudstone: duller-gray-green and more tan than in underlying interval, and thinner- and smoother-bedded. Contains fine-grained mica, comminuted glauconite, and thin beds of impure limestone in upper few feet, and many worm tubes throughout. Brachiopods and *Hyolithes* present but not numerous. Weathers gray-green and forms top of gray-green interval. Base of fossil loc. C9a. *"Amecephalus"* zone	34	10.4
Mudstone: blue-green, banded gray and tan, calcareous, extremely finely arenaceous. Contains oval nodules of bright-tan, limonitic limestone and tiny specks of biotite. Weathers dull-green banded ¼ to 1½ inch drab-tan and forms broken cliffs. Becomes more calcareous upward	59	18.1

	Feet	Meters
Limestone: tan, green, and gray banded, argillaceous, massive fine- to medium-grained, thin-bedded (½ to 2 inches). Weathers buff-tan and forms broken cliffs above large slope. Breaks down to large blocks which weather to platy and fissile fragments	44	13.5
Shale: as in underlying interval but thinner-bedded, more fissile and paler green-gray. Weathers green-tan and forms upper part of slope between cliffs	59	18.1
Shale: green and gray banded, fine-grained, slaty, hard. Contains variable amounts of dark-green poorly developed glauconite. Weathers tan-green banded gray and forms part of slope between cliffs here	53	16.3
Shale and sandstone: dull-gray, fine-grained, thick- and thin-bedded shale and some fissile and chunky finely arenaceous shale very slightly calcareous on joint faces. Glauconite poorly developed in some beds. Occasional intercalated lenses (½ to 1 inch) of extremely calcareous sandstone which contains limonite. Interval weathers dull-gray and green and forms slope here above shale cliffs	50	15.4
Shale and sandstone: dull-green-gray slightly banded, finely arenaceous shale and bands of extremely argillaceous sandstone, all hard, massive and smooth-bedded. A few beds finely micaceous. Some cleavage. Weathers dark-green banded gray	55	16.9
Shale: buff-brown, finely arenaceous, platy to slightly fissile. Slaty cleavage still strongly developed. Weathers buff-brown beneath drab-green interval	41	12.6
Sandstone and shale: deep red-brown (chocolate) and faint-green-black in alternate intervals. Fine-grained, extremely argillaceous and shaly-bedded sandstone or extremely finely arenaceous slaty shale, much sheared. Weathers brown and dark-green-gray in alternate bands and forms steep slope strewn with platy fragments above sheer quartzite cliffs (Pl. 3, fig. 2)	58	17.8
Total thickness of Naiset formation	475	145.9

Note: The section was traced from north spur of Mount Towers westward to Naiset Peak and "tied in" on contact of upper St. Piran massive quartzite against chocolate-colored and black-gray shaly beds at base of Naiset formation. The above part of section was measured on the east side of Naiset Point south and southwest of Gog Lake.

Total thickness measured of Middle Cambrian	732	224.8
Total thickness of Middle Cambrian (measured and estimated)	2925	898.9

LOWER CAMBRIAN

Gog formation (type locality)

Quartzite: white-gray and faint-tan-gray, fine-grained, vitreous, in 2- to 4-foot beds, slightly banded irregularly. Contains *Scolithus* borings in lower beds and numerous ovoid masses of granular tan limonite in isolated areas in upper beds which weather to vesicular cavities ¼ to ⅝ inch in diameter. In lower part occasional ½- to 1-inch bed of pale-gray sandstone. Interval weathers light-tan-gray and forms top of sheer cliffs throughout region	120	36.9

	Feet	Meters
Quartzite: dull-tan-gray, medium coarse-grained, generally thick-bedded. Contains spots of limonite ⅛ inch in diameter and worm borings (*Scolithus*) which extend through some beds 2 feet thick. Near base 1-foot zone of olive-gray nodular sandstone and shale partings. In middle 6- to 8-foot interval of maroon, coarse-grained, thick-bedded, banded quartzite which contains spots of limonite. Interval weathers pink-gray and maroon and forms base of upper sheer cliffs of Gog formation	56	17.2
Sandstone and quartzite: tan and tan-gray, fine-grained, thick- and thin-bedded sandstone. Contains many specks of limonite. Thin beds irregularly rippled and stained green on bedding surfaces. Some thin-bedded intercalated gray quartzite	18	5.5
Note: Section traced south 660 feet to small gulch and continued up the ridge to the top of the St. Piran quartzite.		
Sandstone and shale: dull-green and little maroon, finely arenaceous platy hard shale and gray and tan interbedded platy quartzitic sandstone which contains many specks of limonite. Weathers dark-green-gray and forms bench	18	5.5
Sandstone, quartzite, and shale: dull-gray and tan, fine-grained thin- and flaggy-bedded sandstone, interbedded with thicker-bedded gray quartzite and much finely arenaceous green-tan irregular-bedded shale or argillaceous sandstone which is ripple-marked. Interval weathers conspicuous buff-tan and forms thin flaggy plates	68	20.9
Quartzite: dull-gray, medium-grained, thick-bedded, grading upward into tan-gray finer-grained, thin-bedded (more or less flaggy) quartzite. Weathers tan-gray and forms slope here	111	34.1
Quartzite: pale-gray, fine- to medium-grained, thick-bedded. Weathers gray and forms cliffs. Some thin beds contain *Hyolithes* and worm tubes	94	28.9
Lake Louise shale of Walcott (1928)		
Shale, quartzite, and sandstone: black-gray and extremely dull-green, fine-grained, argillaceous, hard, slaty chunky shale, and in upper part some micaceous shale. Throughout are interbedded dull-tan-gray thin-bedded quartzite and some platy beds of tan-gray calcareous sandstone spotted with dark-tan limonite. Near top a few 2- to 4-inch lenticular beds of very slightly calcareous argillaceous sandstone. Interval weathers dark-tan-gray and forms band in cliffs	62	19.1
Total thickness of "Lake Louise shale"	62	19.1
Quartzite and shale: white-gray, fine-grained, thick- and thin-bedded quartzite. Some beds contain specks of limonite. Weathers light- and tan-gray and forms lower steep slope just west of Wonder Pass. In upper half several 1- to 2-foot intervals of brown, green, and maroon, finely arenaceous shale. Upper 5 to 6 feet of quartzite dark- and pale-gray, medium coarse-grained, and contains much limonite as specks and small flakes	130	40.0
Quartzite and shale: white-gray and some dull-tan, fine-grained, sandy quartzite in 1- to 4½-foot beds and occasional intercalated ½- to 1-foot intervals of dull-tan-green shale. Weathers light-gray and to small angular fragments	66	20.3
Quartzite and shale: faint-tan-gray and maroon, medium- to fine-grained, thick- and thin-bedded quartzite and at top 12 to 14 inches of dull-green and tan, fissile to chunky hard shale	51	15.7

	Feet	Meters
Quartzite: maroon, coarse-grained, less dense and indurated, strongly banded and some cross-bedded, in 8- to 18-inch beds. Weathers maroon	115	35.4
Quartzite: light-gray stained faint-tan grading upward into maroon quartzite	20	6.2
Quartzite: light-pinkish-gray and faint-tan-gray, fine-grained, hard, thick-bedded, occasionally finely banded. Contains a few areas (pebbles?) of milky quartz. Interval covered and forms rounded slope on southwest side of pass	164	50.5
Quartzite: tan, gray, and pink-gray, fine- to medium-grained, slightly vitreous, thick-bedded. Contains much limonite in lower 4 feet and is faintly banded in upper part. Some beds of gray, fine-grained vitreous quartzite in middle part. Weathers faint-tan-gray and pinkish-gray and forms low strike-ridges (hogbacks) here in Wonder Pass	45	13.8
Quartzite: tan-gray and gray, medium- to coarse-grained, generally thick-bedded, occasionally slightly banded in some beds. Contains specks of limonite and veinlets of white quartz. Weathers pale-tan-gray and forms ledges on southeast side of Wonder Pass	23	7.1
Quartzite: dull-red-gray banded darker-red, medium- to coarse-grained, cross- and thick-bedded (12 to 20 inches). Contains specks of hematite and irregular stringers of white massive quartz on joint fissures. Weathers dull-red banded red	13	4.0
Quartzite: faint-tan-gray, medium- to coarse-grained, thick-bedded (2 to 36 inches, average 8 inches). Contains specks of limonite and tiny (⅛ to 1 inch) angular milky pure quartz pebbles. Weathers light-tan-gray and forms top of cliffs in small canyon here	39	12.0
Conglomerate: pebbles pale-gray, medium- to fine-grained, dense, pure, hard quartzite, 1 to 14 inches in diameter, average 4 inches, well-rounded and occasional 12- to 14-inch subangular pebbles. Contains small spots of tan-weathered limonite. Upper 11 feet contains fewer and smaller pebbles of darker-gray, coarse-grained quartzite. Matrix approximately same character of quartzite as pebbles	22	6.8
Total thickness of Gog formation	1235	379.9
Total thickness of Lower Cambrian	1235	379.9
Total thickness of Mount Assiniboine section	5460	1678.8

PRE-CAMBRIAN (BELTIAN)

Hector shale

Shale: black-gray, very fine-grained, argillaceous slaty shale, finely wavy banded lighter-gray. Contains much finely disseminated limonite which causes shale to weather rusty brown. Small spots of yellow-ochre common. Slaty cleavage well developed, strikes N. 8° E. to N. 25° W., dips 83° NW. Some intercalated beds of medium-grained sandstone which contains white quartz crystals. Forms rusty-brown rounded slopes in this region. Some talc developed on cleavage planes.

The erosion surface at the top of the Hector shale is not exposed at the base of the above section measured on Wonder Pass but is perfectly exposed (Pl. 5, fig. 2) on a small isolated ridge at a point approximately a quarter of a mile east of Gog Lake. The following supplementary

section measured at this place serves as a detailed check on the basal part of the larger section and furnishes valuable data concerning the character of the Cambrian—pre-Cambrian erosion surface and the basal Cambrian conglomerate in the Assiniboine area.

	Feet	Meters
Basal part of Gog formation		
Conglomerate or pebbly quartzite: 2 beds: tan-gray, medium- to fine-grained, slightly vitreous, hard, massive quartzite matrix. Contains 2- to 4-inch pebbles which are much less numerous than in lower conglomerate. Weathers pink-gray and forms two ledges here	9	2.8
Quartzite: dull- to dark-gray, medium- to coarse-grained, thick-bedded. Upper surface of one bed ripple-marked, ripples 2 to 3 feet in length	4.5	1.4
Conglomerate: tan- and dull-gray quartzite pebbles as much as 17 inches in long diameter, subangular to rounded, and many more pebbles ½ to 2 inches in diameter cemented in quartzite matrix. Basal bed 18 to 20 feet thick, upper beds 4 to 7 feet thick. Weathers tan-gray stained brown on cliff face	53	16.3
Total thickness of exposed quartzite	66.5	20.5
BELTIAN		
Hector shale		
Shale and sandstone: black, fissile, hard platy shale. Upper bed ranges from ½ to 26 inches in thickness where irregularly cut out by erosion. Several interbedded brown-gray irregular beds 1 to 4 inches thick of medium-grained sandstone. In places basal Cambrian conglomerate rests directly on sandstone, and upper beds of Hector shale are cut out	8–9	2.5–2.8

COMPARISON OF ORIGINAL AND EMENDED ASSINIBOINE SECTIONS

The Cathedral dolomite in the emended section is equivalent to the combined Cathedral and Ptarmigan formations in the original section (Fig. 3). The Naiset formation is overlain by dolomite in some parts of the area and by limestone in others. For example, on Mount Towers the basal part of the Cathedral is a buff-gray-weathering massive-bedded sugary unfossiliferous dolomite; but on Naiset Point (Pl. 5, fig. 1), and apparently on Wedgwood Peak, the basal Cathedral consists of black-gray thin- and some thick-bedded fossiliferous limestone which weathers dark-gray. On Ptarmigan Peak, Castle Mountain, and Mount Bosworth—the other sections studied by the writer in which the Ptarmigan limestone was precipitated (Deiss, 1939a, p. 1002-1004)—it is 200 to 385 feet below the *Albertella* zone in the Cathedral dolomite. Many individuals of several species of *Albertella* occur in the basal limestones of the Cathedral formation on Naiset Point from 148 to 257 feet above the top of the Naiset formation, which in its upper 56 feet contains species of *Kochaspis* and *Nisusia*. These trilobites are associated in the Ptarmigan limestone on Ptarmigan Peak. Consequently, the limestones which rest on the

Naiset formation in the Assiniboine area are approximately of the same age as the lower-middle Cathedral dolomite on Mount Bosworth (the type locality), do not constitute a mappable group of rocks even within

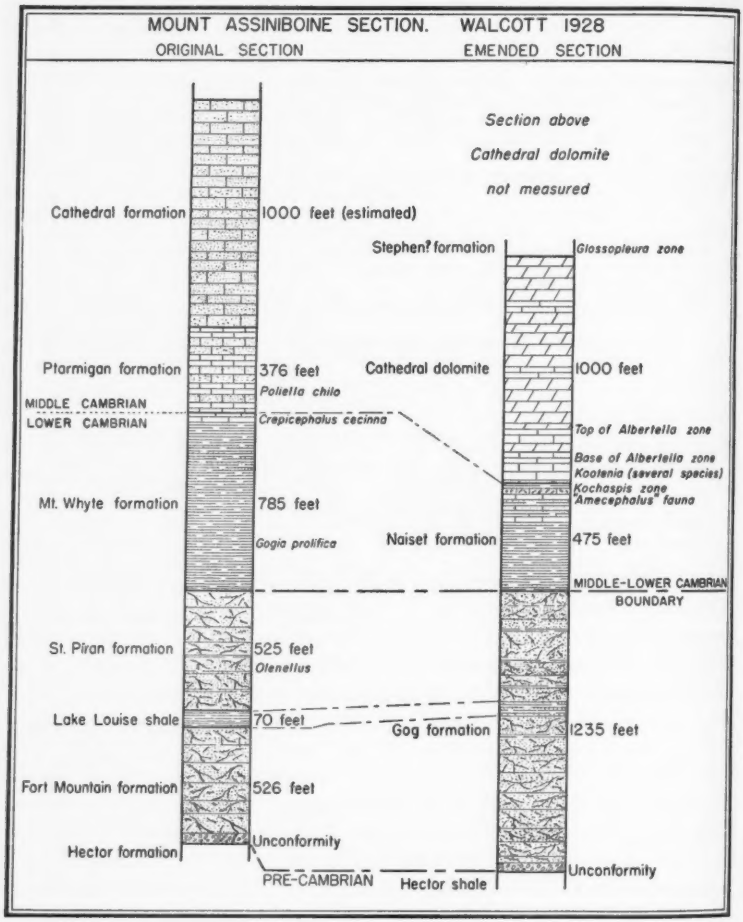

FIGURE 3.—*Comparison of original and emended Mount Assiniboine sections*

all parts of the Assiniboine region, and cannot be assigned to the Ptarmigan formation. These limestones do contain a species of *Poliella*, which may be *P. chilo* (Walcott, 1928, p. 297), but this species is typical of

the *Albertella* fauna on Ross Lake and Mount Bosworth and only substantiates the assignment of these limestones to the Cathedral formation.

The rocks between the top of the Gog and the base of the Cathedral formations on Mount Towers, Naiset Point, and Wedgwood Peak are not more than 500 feet thick and total only 475 feet on Naiset Point (Fig. 3). Walcott gave 785 feet for the thickness of these rocks and assigned them to the Lower Cambrian Mount Whyte formation (Walcott, 1928, p. 297-298). In addition to the fact that these sediments are only two-thirds as thick as originally stated they also contain *"Amecephalus", Kochaspis,* and other Middle Cambrian fossils in the upper 56 feet. (*See* faunal list under "Naiset formation" in this paper.) Unfortunately fossils could not be found in the middle or lower parts of the formation. However, the formation is lithologically homogeneous and consists of indurated finely arenaceous and calcareous shales interbedded with occasional argillaceous sandstones in the lower part and argillaceous limestones and thin intraformational conglomerates in the upper part. The formation probably was deposited continuously during one epoch, and the only known fossils are Middle Cambrian in age. Therefore, the Naiset formation is tentatively assigned to the Albertan. The Middle Cambrian age of the formation is more probable because the only undoubted Lower Cambrian fossils ever reported from the Assiniboine area were fragmentary heads of *Olenellus* which Walcott (1928, p. 298) said were present in the lower 290 feet of the "St. Piran", at least 235 feet below the base of the Naiset formation. The rocks assigned to the Mount Whyte formation in the original Mount Assiniboine section, therefore, are placed in the new formation—the Naiset—and the Middle-Lower Cambrian boundary is tentatively drawn at the top of the underlying quartzitic sandstones (Fig. 3).

The basal clastic sediments in the Assiniboine section are 1235 feet instead of 1121 feet thick as originally stated and are divisible into three lithologic units which closely resemble the Fort Mountain, Lake Louise, and St. Piran clastics at the north end of the Bow Range and in the vicinity of Ptarmigan Peak and Redoubt Mountain (Fig. 1). Near Mount Assiniboine, as in the northern region, these quartzitic sandstones and shales generally form steep slopes and sheer cliffs and consequently form narrow or small areas on a geologic map. The Lake Louise shale was found to be too thin to map on a scale of 1/62,500 in the northern area and for that reason was made a member of the Fort Mountain sandstone (Deiss, 1939a, p. 997). In the Mount Assiniboine area the shale which Walcott called Lake Louise is not only too thin to map but is usually covered with soil or talus to the extent that the quartzitic sandstones above and below it are not divided by any observable boundary

and consequently are not mappable as separate formations. Further, the Lake Louise shale occurs at a different stratigraphic horizon above the base of the Cambrian or beneath the Mount Whyte-St. Piran boundary in every section in which the shale has been examined by the writer and may not even be a persistent member of the Fort Mountain in the restricted northern area. In the Assiniboine area the absence of the Mount Whyte formation, the presence of the quartzites immediately beneath the Middle Cambrian, and the occurrence of *Olenellus* 235 to 500 feet below the top of the quartzites whereas it is in the basal Mount Whyte on Ptarmigan Peak, Mount Bosworth, Ross Lake, and other sections in the north, all strongly suggest that the entire clastic series at the base of the Cambrian in the Assiniboine region may be equivalent to the Mount Whyte formation and possibly to part of the St. Piran in the northern sections. If this be true the 62-foot shale interval in the Gog formation at Mount Towers, Naiset Point, and Mount Wedgwood is merely another shale phase in the clastic series at the base of the Cambrian and bears no relationship to the Lake Louise shale. Because of these facts the basal Cambrian quartzites, sandstones, and shale are all assigned to the new Gog formation in the emended section (Fig. 3), and the names St. Piran, Lake Louise, and Fort Mountain are discarded for use in the Assiniboine area.

CAMBRIAN-BELTIAN EROSION SURFACE IN MOUNT ASSINIBOINE AREA

Walcott (1910, p. 423-431) was the first to describe the Beltian (pre-Cambrian) rocks of the Canadian Rockies. As a result of his work in the Bow Valley and adjacent ranges Walcott described the pre-Cambrian rocks and assigned them to two formations—the Hector shale and the underlying Corral Creek quartzitic sandstone. He correlated them with the Kintla and Sheppard formations (Willis, 1902, p. 324) in Glacier Park and with the Camp Creek formation in the Swan Range of northwestern Montana (Walcott, 1906, p. 18). In the Bow Valley area every known exposure of the Cambrian—pre-Cambrian contact exhibits the basal pebbly Cambrian sandstones resting almost conformably on the slightly eroded upper surface of the Hector shale, but nowhere in the area are the rocks of the two systems as sharply demarcated as on the north side of the Assiniboine "Massif" (Pl. 5, fig. 2). Walcott (1928, Pl. 60) gave two photographs showing the basal Cambrian conglomerate and the Cambrian-Beltian contact but did not discuss the unconformity. A maximum thickness of 200 to 250 feet of Beltian shales is exposed which may be the southern phase of the Hector.

The Cambrian-Beltian relationships on the north side of the Assiniboine "Massif" are of special interest and importance. The best ex-

posures of the Cambrian—pre-Cambrian contact yet seen by the writer in the Canadian Rockies are on the northeast side of Naiset Point (Pl. 5, fig. 2) on the small hill which is approximately three-sixteenths of a mile east of Gog Lake and at the north end of the spur of Wedgwood Peak. Another excellent exposure is on the southeast shore of Magog Lake on the short blunt peninsula situated approximately three-quarters of a mile from the south end of the lake. At the locality northeast of Naiset Point the basal Cambrian conglomerate consists of dull- and tan-gray medium- to coarse-grained quartzite pebbles as much as 17 inches in long diameter (average 2 inches) embedded in tan-gray medium- to fine-grained slightly vitreous quartzite. Most of the pebbles are well rounded (Pl. 5, fig. 2), and a few are subangular. The basal bed is 18 to 20 feet thick, and the upper part of the conglomerate is in beds 4 to 7 feet thick. The total thickness of the conglomerate at this place is at least 66 feet. In the emended section on Wonder Pass (Pl. 4, fig. 2) the basal conglomerate totals only 22 feet in thickness, and, although the maximum size of the pebbles is not great, the average (4 inches) is larger than that near Gog Lake.

On the north end of Wedgwood Peak at a point approximately 2¼ miles west of the outcrop at Naiset Point the conglomerate is again only 20 to 30 feet thick, but the matrix is more sandy, the pebbles are fewer and much smaller, and the basal bed is 1 to 3 feet thick. Thus the conglomerate thickens 40 feet or more toward the northeast side of Naiset Point both from the east and west.

The large pebbles in the conglomerate are also larger than any observed elsewhere in the basal Cambrian of the Cordilleran region in the northern United States or in the Canadian Rockies. The large pebbles imply that the north shore of the Montana Island was not far to the south as wave action or marine currents could not have transported such pebbles long distances. The rounded shape of the pebbles, their unusually large size in the vicinity of Naiset Point (Fig. 1), and the fact that the conglomerate thins rapidly to the east and west all suggest that the east side of Naiset Point may mark the channel of an old stream which flowed northward and debouched its load of gravel at that point.

In the Lake Louise region the basal Cambrian beds are more nearly pebbly sandstones than conglomerates. The pebbles range in diameter from $\frac{1}{16}$ to 1¼ inches and average a quarter of an inch, and the underlying Hector is somewhat similar lithologically to the shale in the lower Fort Mountain formation. The Hector also contains intraformational conglomerates and quartz pebbles similar to those in the overlying Cambrian sandstones which seem to rest almost if not quite conformably upon

the Hector. These facts caused some doubt as to the pre-Cambrian age of the Hector in the vicinity of Ptarmigan Peak and the Bow Valley (Deiss, 1939a, p. 1012). The pre-Cambrian shale in the Assiniboine area is strikingly different in composition from the overlying Gog formation, and the two are separated by an irregular wavy erosion surface which cuts out 2 feet of beds within a horizontal distance of 15 to 20 feet (Pl. 5, fig. 2). Therefore the Assiniboine area furnishes conclusive evidence that the Hector shale and underlying Corral Creek formation are Beltian as originally stated by Walcott (1910, p. 427-430).

The Beltian sediments in the Assiniboine area formed part of the Montana Island during the Cambrian-Algonkian interval (Deiss, 1935, p. 106) and probably were eroded longer than those in the Bow Valley because the Gog formation is younger than the Fort Mountain sandstone. Consequently erosion continued in the Assiniboine area while the Fort Mountain and possibly the lower part of the St. Piran were being deposited to the north. In any event the Cambrian is more clearly and sharply separated from the Beltian in the Assiniboine area than in any other place seen by the writer in the Canadian Rockies.

CAMBRIAN FORMATIONS

GENERAL STATEMENT

The Lower and Middle Cambrian formations in the vicinity of Castle Mountain, Ptarmigan Peak, and Ross Lake were emended in 1939 (Deiss, 1939a, p. 993-1008) with the exception of the St. Piran and Stephen. In addition the new Upper Cambrian Pika formation was defined. The field work upon which that revision was based was controlled by the principle that a formation must be a mappable natural lithologic unit if it is to have any value in deciphering the stratigraphic, structural, and economic geology of the area (Deiss, 1939a, p. 993-994). This principle and the others formulated by the committee of the Geological Society of America on classification and nomenclature of rock units (Ashley, *et al.*, 1933, p. 423-447) have also been the criteria upon which are defined the new formations given in this paper. Because Mount Bosworth is in the same area as the sections remeasured in 1938 and because the present report and the field work upon which it is based are a continuation of work done in 1938 this paper is a companion to that published in 1939 (Deiss, 1939a, p. 951-1026), and the two should be used together.

Of the Lower and Middle Cambrian formations revised or discussed in 1939, only the Ptarmigan and Stephen need additional discussion at this time. Two new formations—the Gog and Naiset—are defined.

GOG SANDSTONE

General statement.—In the Mount Assiniboine region the lithologic sequence appears superficially to be the same as in the Ptarmigan Peak area. The basal Cambrian rocks are conglomerate, quartzitic and pure sandstone, and intercalated arenaceous shale. These rocks are overlain by shales, argillaceous thin-bedded sandstone, and impure limestone which in turn are overlain by either massive-bedded tan-gray sugary dolomite or black-gray thin-bedded fine-grained limestone. This sequence simulates respectively the Fort Mountain and St. Piran, Mount Whyte, and Cathedral or Ptarmigan formations on Ptarmigan Peak. Consequently Walcott (1928, p. 297-298) assigned these rocks in the Assiniboine region to these formations.

In 1939 the upper part of the rocks which Walcott called "Mount Whyte" on Naiset Point was found to contain *Kochaspis chares* (Walcott), *"Amecephalus" cleora* (Walcott) and other basal Middle Cambrian trilobites. Fossils could not be found in the middle or lower parts of the formation, but, as it is nearly lithologically homogeneous and no stratigraphic break could be found, sedimentation was probably continuous. On the other hand Walcott (1928, p. 298) reported *Olenellus* from the beds somewhere between 335 and 525 feet below the base of his Mount Whyte in the Assiniboine section. The writer has never been able to find *Olenellus* anywhere except in the basal part of the Mount Whyte formation in the vicinity of Bow Valley, although the trilobite has been found in the upper quartzitic sandstones of the St. Piran formation near Lake Louise.

These facts show conclusively that the rocks in the Assiniboine section formerly assigned to the Mount Whyte are basal Middle Cambrian and that those assigned to the St. Piran may be Lower Cambrian and probably are of the same age as the Mount Whyte at its type locality on Ptarmigan Peak (Deiss, 1939a, p. 998-1000). Consequently the rocks which Walcott assigned to the Lake Louise and Fort Mountain formations in his Assiniboine section are younger than those formations and more probably are of the same age as the St. Piran quartzite, and, therefore, the clastic rocks at the base of the Assiniboine section must be defined as a new formation.

The lower 1235 feet of Cambrian rocks in the Assiniboine region consists of arenaceous and pure, thick- and thin-bedded, unfossiliferous quartzites and sandstones. At the base is a variable thickness of conglomerate, and approximately 688 feet above the base is a 62-foot interval of arenaceous shale and argillaceous sandstone. This clastic series is not separable into mappable units and consequently must be combined into one formation. These rocks constitute the new Gog forma-

tion, the type locality and section of which is the crest of Wonder Pass and the lower part of the northeast spur of Mount Towers. The type section thus lies in the NE. ¼ sec. 21, and NW. ¼ sec. 22, T. 22 N., R. 12 W., British Columbia (Pl. 4, fig. 2).

Original definition.—The Gog formation takes its name from Gog Lake at the northeast edge of Naiset Point. The lake basin is cut in the quartzites and sandstones of the Gog formation which forms sheer cliffs and steep slopes on the north side of the Assiniboine "Massif" at the base of Mount Towers, Naiset Point, Mount Magog, Mount Wedgwood, and Mount Watson.

The lower 20 to 66 feet of the Gog formation consists of thick-bedded quartzite-conglomerate which varies greatly in thickness and lithologic character within short distances. At the type locality on the southeast side of Wonder Pass the conglomerate is 22 feet thick and consists of pale-gray stained tan, medium- to fine-grained, dense pure quartzite pebbles 1 to 14 inches in diameter, cemented in a matrix of gray, medium-grained quartzite. The pebbles are well rounded to subangular in the lower beds and are fewer and smaller in the upper beds. In the vicinity of Gog Lake (Pl. 5, fig. 2) the pebbles attain their maximum size of 17 inches in long diameter, and the conglomerate is much thicker bedded. This basal conglomerate of the Gog formation is better developed in the Assiniboine area than in any other place where the Cambrian rests upon Beltian rocks.

Above the conglomerate is 666 feet of light-gray, pinkish, tan, and dull-maroon, interbedded pure and sandy quartzites in beds which range from 1/6 to 4 feet and average 1 foot in thickness. The quartzites are strongly cross-bedded and banded locally and contain many specks and small flakes of limonite irregularly distributed. Small (1/8- to 1-inch) subangular milky pure quartz pebbles are numerous in the lower beds and are irregularly scattered in the higher beds of this interval. The quartzites weather variably drab-gray, tan, and maroon. In the upper 240 feet of this interval are occasional 1- to 2-foot intercalated beds of green, brown, and maroon, finely arenaceous, chunky to fissile shale. The rocks in this interval and the underlying conglomerate are those which Walcott (1928, p. 298) called Fort Mountain in his Assiniboine section.

In the upper-medial part of the Gog formation is a 62-foot interval of black-gray and drab-green, fine-grained, hard, slaty to micaceous shale which weathers dark-tan-gray and forms a conspicuous band in the cliffs. Platy quartzite and calcareous sandstone are interbedded through-

FIGURE 1. CHARACTERISTIC NAISET-GOG CONTACT
As exposed in cliff faces. West side of Mount Towers from Naiset Point. N = Lower Part of Naiset formation, G = Gog sandstone, W = Wonder Peak.

FIGURE 2. TYPE SECTION OF GOG SANDSTONE
On east side of Mount Towers. Wonder Pass on right. N = Lower part of Naiset formation, G = Gog sandstone, H = Hector (pre-Cambrian) shale, C = Cathedral limestone.

NORTH RIDGE OF MOUNT TOWERS

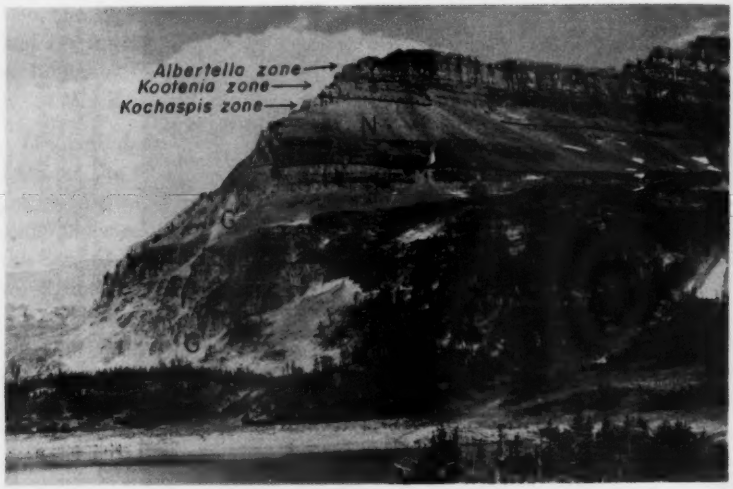

FIGURE 1. WEST SIDE OF NAISET POINT ABOVE MAGOG LAKE
C = Lower Cathedral limestone (light areas dolomite), N = Naiset formation, G = Gog sandstone.

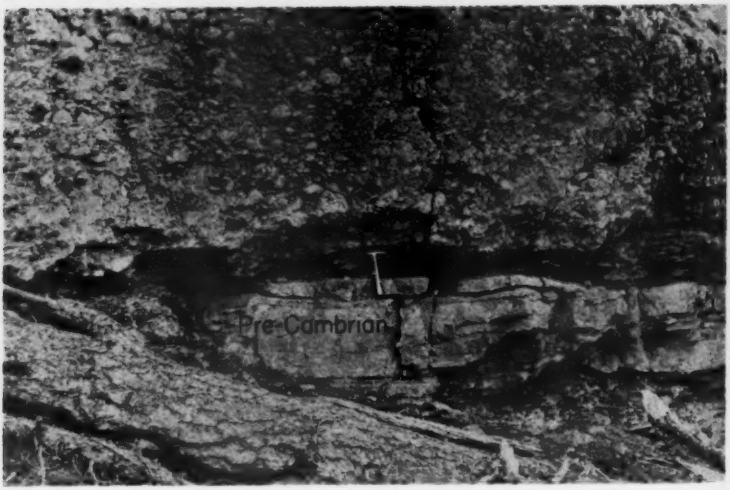

FIGURE 2. BASAL CAMBRIAN CONGLOMERATE AND PRE-CAMBRIAN EROSION SURFACE
East of Gog Lake. Length of hammer handle 15 inches.

NAISET POINT AND PRE-CAMBRIAN EROSION SURFACE

out the shale. This is the interval which Walcott referred to the Lake Louise shale.

The upper 485 feet of the Gog formation consists of massive-bedded tan and gray sandstones and quartzites and some intercalated shales. Occasional beds contain specks and larger irregular blebs of limonite, and in the upper 175 feet the thicker beds are marked with numerous *Scolithus* borings. The rocks weather tan-gray and form the upper light-colored cliffs below the darker slopes of the Naiset shales and sandstones (Pl. 4, fig. 1). Walcott (1928, p. 298) reported fragments of *Olenellus* from beds somewhere between 235 and 525 feet below the top of the Gog formation. These fossils could not be found by the writer in the field, and a collection from these beds is unknown in the United States National Museum. The only other fossils found in the Gog formation were a few poorly preserved *Hyolithes* casts, and worm tubes from a horizon approximately 450 feet below the top of the formation.

NAISET FORMATION

General statement.—The rocks in the Assiniboine region between the Gog and Cathedral formations contain Middle Cambrian fossils in the upper 56 feet, are unfossiliferous throughout most of the formation, and differ lithologically from any other formation to the north in the vicinity of Ptarmigan Peak or Mount Bosworth (Fig. 1). Walcott (1928, p. 297-298) gave 785 feet as the thickness of these rocks and assigned them to the Lower Cambrian Mount Whyte formation. The reasons why these rocks cannot be correlated with the Mount Whyte and must be defined as a new formation are given in the "General statement" under the Gog formation. They are here described as the Naiset formation and are believed to have been deposited during the older part of the Albertan epoch. The horizontal extent and variation in thickness of the formation are unknown, but the rocks constituting it are well exposed and uniform in thickness throughout the northern part of the Assiniboine area.

The type locality and section of the Naiset formation is on the upper-middle part of the east side of Naiset Point (Pl. 3, fig. 2) in the NW. ¼ sec. 21, and the SW. ¼ sec. 28, T. 22 N., R. 12 W., British Columbia.

Original definition.—The formation, named from Naiset Point, has a total thickness of 475 feet on that mountain and is divisible into four lithologic members of which the upper one is clearly demarcated from the others. The formation is sharply marked at the base by the contact of the dark shales against the light-colored quartzite which forms the top of the Gog formation (Pl. 3, fig. 2), and at the top by the drab- to

bright-tan shales and limestones beneath the black-gray limestone or pale-gray dolomite which form the base of the Cathedral formation.

The lower member of the Naiset formation is 58 feet thick in the type locality and consists of deep-red-brown (chocolate-colored) and pale-green-black shale in alternating bands and fine-grained extremely argillaceous shaly-bedded sandstone. These beds weather brown and dark-green-gray in alternating bands and form dark steep slopes above the sheer quartzite cliffs of the Gog formation (Pl. 4, fig. 1).

The second member or interval consists of 258 feet of dull-green-gray, dull-gray, and buff-brown, fissile and chunky, more or less banded, hard, finely micaceous shale which contains poorly developed glauconite on the surfaces of some beds. Throughout the interval are irregularly interbedded thin lenses (½ to 1 inch) and beds of argillaceous, calcareous, limonitic sandstone. The rocks weather dull-green in contrast to the darker basal beds.

The overlying interval is 137 feet thick and consists of blue-green, tan, and gray, banded mudstones, and tan and gray fissile argillaceous limestones. The mudstones contain finely comminuted mica flakes, oval nodules of bright-tan limonitic limestone, and weather dull-green banded drab-tan ¼ to 1½ inches. Cranidia of *"Amecephalus,"* shells of *Hyolithes,* and several genera of brachiopods are present in the beds in the upper 34 feet of this interval.

The upper 22 feet of the Naiset formation is more variable lithologically than the remainder of the formation and contains most of the fossils found within it, and were it not for the fact that the same species of *"Amecephalus"* occur at the very top of this interval as in the underlying 34 feet but are not present in the overlying limestones these beds might have been placed at the base of the Cathedral formation. The contrast in color between them and the underlying beds is just as great as that with the overlying limestones. The lower third of this interval consists of bright-tan finely arenaceous and calcareous shale in beds 1/64 to ⅛ of an inch in thickness. Intercalated in the bright-tan-weathering shale are 2- to 11-inch beds of orange-tan fossiliferous intraformational conglomerate which contains tiny green limestone pebbles and much limonite. The upper two-thirds of the interval consists of dull-gray, fine-grained, argillaceous limestone finely banded tan. These rocks weather drab-tan-gray beneath the black-gray basal Cathedral limestone on Naiset Point. On Mount Towers these beds are overlain by pale-buff-gray massive bedded sugary dolomite, and the basal black limestones are absent.

The fossils in the following faunal list were collected on Naiset Point (Loc. C9a) in the type section from the upper 56 feet of the formation.

When they were collected they were separated into three zones, but further study in the laboratory showed that the same species range throughout the 56 feet of beds. The list of genera and species given here must be considered tentative until the fauna can be more carefully studied and the new species described.

"Amecephalus" cleora (Walcott)
"A." 3 n. sp.
n. gen. cf. *Amecephalus* Walcott
Kochaspis n. sp. cf. *K. chares* (Walcott)
K. n. sp. cf. *K. gogensis* (Walcott)
n. gen. cf. *Kochiella Poulsen*
Poliella? sp. indet.
P.? n. sp.
n. gen. of trilobites
Nisusia? 2 n. sp.
Wimanella? sp. indet.
Hyolithes sp.

PTARMIGAN LIMESTONE

General statement.—The Ptarmigan limestone and Cathedral dolomite are closely related in origin. The history of the Ptarmigan was discussed at length in 1939 (Deiss, 1939a, p. 1000-1001) when the writer pointed out that:

"The Ptarmigan limestone has been the most confused and misunderstood Middle Cambrian formation in the Canadian Rocky Mountains. The unfortunate and incorrect assignment of the "*Albertella* fauna" to the Ptarmigan has caused probably more erroneous statements and fallacious conclusions during the past 20 years concerning the fossils and correlation of this formation than any other in the Cordilleran trough.

". . . Apparently, the formation was set up solely to provide a name for the strata in which the *Albertella* fauna was believed to occur, but Burling (1916, p. 469-472) had previously shown that the *Albertella* fauna was several hundred feet above the base and well within the Cathedral formation. *Albertella* and its associated genera are not present in the type locality of the Ptarmigan formation, where fossils are extremely rare in the formation, and the few which are present occur only in the lower and upper parts. What is even more peculiar, the rocks assigned to the Ptarmigan in the Ross Lake section (Walcott, 1917b, p. 14-15) are tan-gray and buff dolomites typical of the Cathedral formation throughout the area."

The field work in 1938 showed that the Ptarmigan limestone is a readily mappable unit in the vicinity of Ptarmigan Peak and Redoubt Mountain (Deiss, 1939a, Pl. 6; Pl. 7, fig. 1) but that on Castle Mountain the formation has thinned to less than one-fourth its thickness, is not represented in the Ross Lake section, and does not contain *Albertella* or other genera associated with it. Because the Ptarmigan is sharply demarcated and readily mappable near Ptarmigan Peak the writer emended the formation (Deiss, 1939a, p. 1000-1001) and accepted Walcott's (1917a, p. 3-4) suggestion that:

"My first field impression was that the Ptarmigan limestone was deposited locally in a shallow basin largely as oolites before the coming of the physical change that produced the great Cathedral limestones."

The field work in 1939, however, threw much new light on the problem and suggested a quite different cause for the origin of the Ptarmigan limestone. Throughout the sections remeasured and examined in Banff and Yoho National Parks and especially in the Assiniboine region the Cathedral formation consists of massive-bedded, fine to coarsely crystal-

line sugary dolomite which is pale-tan-gray or pure-light-gray on fresh fracture and weathers light-gray or mottled light- and buff-gray. Isolated irregularly shaped masses of sooty-black and lead-gray, fine- to medium-grained, thin-bedded, locally oölitic limestone occur throughout the Cathedral but are more numerous and larger in the lower part of the formation. These limestone masses weather dark-gray in contrast to the surrounding or overlying dolomite and range from a few to 460 feet in thickness and from less than 100 feet to 20 miles or more in horizontal extent. Fossils have never been found in the dolomite, but the limestone carries rare and poorly preserved trilobites and brachiopods. On Ptarmigan Peak and elsewhere the Cathedral contains these isolated stringers and irregular areas of black limestone well up in the dolomite (Deiss, 1939a, p. 979). The opposite condition prevails less commonly but is well developed on Naiset Point where irregular masses of dolomite 8 to 18 feet thick and more than 100 feet in diameter are surrounded by black fine-grained limestone. This fact seems to substantiate the correctness of Swartzlow's (1933, p. 331) conclusion that such dolomite and limestone are syngenetic. The presence of isolated masses of dolomite such as those on Naiset Point 175 feet below the top and at least 60 feet above the base of the limestone precludes the possibility that the dolomite represents later alteration of the limestone.

In his hypothesis of possible contemporaneous deposition of dolomite and limestone in the Chouteau formation of Missouri Swartzlow (1933, p. 336) states:

> "Another possible explanation . . . is that both calcite and dolomite were precipitated at the same time. Growing aggregates of dolomite would attract the double salt and aggregates of calcite would attract the calcium carbonate. As the calcium carbonate in the sea-water was becoming depleted, the precipitation of dolomite would dominate until it became the only salt that was being deposited."

This hypothesis gives a rational explanation for the preponderance of limestone masses in the base or lower part of the Cathedral dolomite in Alberta and British Columbia.

In light of the field relationships outlined above, the suggestion is tentatively offered that the Ptarmigan formation represents local deposition of 460 feet of limestone in the immediate vicinity of Ptarmigan Peak, 145 feet at Castle Mountain, and 195 feet at Mount Bosworth (Fig. 4) while dolomite was being deposited at Ross Lake and elsewhere. This suggestion also explains the abrupt disappearance and apparently haphazard thinning of the Ptarmigan limestone within short distances.

The Ptarmigan limestone in its type locality on Ptarmigan Peak contains the basal Middle Cambrian *Kochaspis* fauna. The Cathedral dolomite on Castle Mountain and Mount Bosworth contains the Ross Lake shale with its *Albertella* fauna 200 to 385 feet respectively above the base

of the formation. On Naiset Point, 50 miles southeast, the *Kochaspis* (Ptarmigan) fauna is in the upper beds of the Naiset formation, and *Albertella* is in the 257 feet of limestone which forms the basal part of the Cathedral formation. Obviously, therefore, the rocks which Walcott (1928, p. 297) assigned to the Ptarmigan in his Assiniboine section are almost precisely the same age as those 200 and 385 feet above the base of the Cathedral on Castle Mountain and Mount Bosworth.

These facts show clearly that the Ptarmigan is not a meaningful or useful formation except on Ptarmigan Peak, Redoubt Mountain, and possibly on Castle Mountain; that the age of the black-gray limestones if and when they are present at the base of the Cathedral formation south of Castle Mountain should not be called Ptarmigan or correlated with that formation unless they contain the *Kochaspis* fauna; and that such limestones are present and irregularly developed in widely separated localities at the base of the older Middle Cambrian dolomite formation.

Possibly it would have been better in 1939 (Deiss, 1939a, p. 1001-1002) to have rejected the Ptarmigan as a separate formation and included the limestones on Ptarmigan Peak and Castle Mountain in the Cathedral formation. However, such a procedure would have caused much confusion in the minds of stratigraphers and paleontologists unfamiliar with the field relationships. In the future the Ptarmigan should be thought of as the oldest Middle Cambrian formation in the northern part of the area but one that is readily mappable only in the vicinity of Ptarmigan Peak and Redoubt Mountain.

STEPHEN FORMATION

Walcott (1908a, p. 3-4) first named the Stephen formation and placed its type section on Mount Stephen. In doing so he stated that the Stephen is "562 feet thick, with 150 feet of local development of *Ogygopsis* shales at the summit." The same year Walcott (1908c, p. 236-239) published a detailed section of the formation on Mount Stephen in which he gave the thickness of the formation as 712 feet and thus clarified his earlier statement. In 1928 Walcott (1928, p. 315-319) again published the Mount Stephen section but now placed the 190 feet of "Bluish gray limestone with bands of dark siliceous shale in lower portion" in the upper part of the Stephen instead of in the base of the Eldon and gave for this 190-foot interval a faunal list which included *Ogygopsis* and other trilobites that also occur in the underlying *Ogygopsis* shale. In the same paper he gave the following revised definition of the formation (Walcott, 1928, p. 247):

> "Character.—On Mount Stephen, calcareous, siliceous, and finely arenaceous shales with more or less argillaceous matter, superjacent to a thick band of bluish-black limestones in thin layers, which in turn is underlain by a series of oölitic and gray limestone alternating with bands of dolomitic limestone and a few bands of siliceous shale."

Clearly he again stated that the *Ogygopsis* shale is at the top of the Stephen formation.[7]

The late Albertan age of the *Ogygopsis* shale fauna is indicated by such trilobites as *Agnostus montis* Matthew, *Bathyuriscus rotundatus* (Rominger), and *Elrathina cordillerae* (Rominger). This assemblage of genera is restricted to one horizon in the upper part of the Middle Cambrian elsewhere in the Cordilleran region. A closely related fauna is known from many similar species in the Pentagon shale (Deiss, 1939b, p. 42-44) and Meagher limestone (Deiss, 1936, p. 1330-1332) of Montana. However, the *Ogygopsis* fauna has never been found elsewhere in Alberta and British Columbia. Examination of the published faunal lists of the Mount Stephen section shows that nearly all the other trilobites in the formation are likewise more characteristic of late than of middle or early Albertan time. In fact, Walcott (1908c, p. 238) states that *Agnostus* and *Bathyuriscus* are present at least 240 feet above the base of the Stephen on Mount Stephen.

In contrast, the "Stephen formation" on Castle Mountain, Ptarmigan Peak, Ross Lake, and Mount Bosworth contains a prolific *Glossopleura* fauna at its base (Deiss, 1939a, p. 1005-1006). This fauna is one of the most widespread in the Cordilleran trough and is everywhere limited stratigraphically to a narrow zone. Strangely, *Glossopleura* has never been reported from Mount Stephen, and the writer could not find it there in 1939. The fauna is well represented on Mount Bosworth 6½ miles northeast of the outcrop of *Ogygopsis* shale on Mount Stephen. On the other hand *Thomsonaspis* occurs in the upper beds of the "Stephen" on Castle Mountain (Deiss, 1939a, p. 967) but has not been found on Mount Stephen 25 miles to the northwest. The *Thomsonaspis* fauna is the youngest Albertan biota yet published from the Cordilleran trough, occurs nearly 400 feet above the *Bathyuriscus-Elrathina* fauna in northwestern Montana, and, therefore, is younger than the *Ogygopsis* fauna.

When the Mount Stephen section was examined in 1939 several surprising facts were discovered. The *Ogygopsis* fauna occurs in several different kinds of matrix: brown soft platy calcareous finely arenaceous shale in the upper beds; green tan platy harder shale in the underlying beds; black-gray calcareous shale beneath the green and tan interval; and white-buff soft slightly calcareous clay-shale at the base. The shale strikes N. 10° to 20° W., and dips 35° to 42° SW. and more steeply than the slope of Mount Stephen at that place. Consequently, as one ascends the slope he goes down-section through the *Ogygopsis* shale. At an ele-

[7] The Burgess shale between Mount Field and Mount Wapta (Walcott, 1912b?, p. 149-151) does not enter into the present problem and must not be confused with the *Ogygopsis* shale which is known only from Mount Stephen.

vation of approximately 7000 feet the *Ogygopsis* shale rests upon steel-gray, buff-weathering, slightly mottled, thin-bedded Cathedral dolomites instead of being overlain by the Eldon dolomite as had been reported. At 7200 feet in elevation the Cathedral dolomite is faulted against what appears to be thicker-bedded dolomites of the basal Eldon formation.

The foregoing facts suggest that on Mount Stephen the *Ogygopsis* shale forms the base instead of the upper part of the Stephen formation, that the published type section is largely inaccurate and possibly inverted, and that beds equivalent to the *Glossopleura* zone on Castle Mountain, Ptarmigan Peak, and elsewhere are absent. If this be true the Stephen in the type locality is partly different lithologically and in age than the rocks assigned to the Stephen formation in the sections to the northeast and southeast, and a hiatus is present at the base of the formation which is equivalent to at least the time required for the deposition of part or all of the beds containing *Ogygopsis* on Mount Stephen.

In summary, the Stephen formation remains a partly unsolved problem and should not be emended until much more field work is completed. The formation is of a different age in the other sections than on Mount Stephen which was prematurely chosen as its type locality. The limestones and shales which separate the massive Cathedral and Eldon dolomites throughout the area will probably always be mapped as a separate formation and may be called the Stephen because of their comparable stratigraphic position. In age these rocks range from *Glossopleura* to *Thomsonaspis* time and are equivalent to the 900 to 1200 feet of beds comprising the Damnation, Dearborn, Pagoda, Pentagon, and part of the Steamboat formations (Deiss, 1939b, p. 38-45) in Montana. Conclusive evidence for the solution of the Stephen problem will be difficult to obtain because the rocks on Mount Stephen are much deformed and faulted and the *Ogygopsis* shale may not be present in the unexplored adjacent region to the east.

STRATIGRAPHIC CORRELATIONS

GENERAL STATEMENT

During the last 7 years' field work on the Middle and Lower Cambrian formations in the Cordilleran trough, 37 stratigraphic sections were measured at various points from southeastern Nevada to central-western Alberta. These sections were deliberately chosen because they were well exposed and contained strategically located faunal zones. Throughout the field and laboratory studies confusion always has arisen when correlations were attempted between formations in widely separated regions. The confusion results because formations must be more or less arbitrary lithic units which generally are mappable only in small areas. That strata

and, therefore, formations do not usually have great horizontal dimensions has long been known, but even experienced field geologists find difficulty in visualizing the extent to which lithic units (formations) overlap and interfinger in relatively short distances. In addition fossils, which are generally accepted as the ultimate and most precise basis of correlation between widely separated sections, unfortunately, are only time markers and in many instances bear no constant relationship to a lithic mappable unit or even to one type of rock. For example, such an assemblage as *Albertella helena, "Amecephalus" stator, Kochina bosworthensis, Poliella chilo,* and *Ptarmigania rossensis?* occur in a 2-inch soft fissile green shale on Castle Mountain, in a 7-foot black-green hard siliceous slaty shale on Mount Bosworth and near Ross Lake, and in a 100-foot interval of black fine-grained massive-bedded limestone on Naiset Point. In Montana, Walcott (1917a, p. 17) reported *Albertella* from the green soft micaceous Gordon shale. In Two Mile Canyon, Idaho, the writer found *Albertella* in a massive gray medium- to coarse-grained limestone, and in the Highland Range of Nevada *Albertella* occurs in a similar limestone (Deiss, 1938b, p. 1153) in the Comet shale. Obviously, the beds which contain *Albertella* are of approximately the same age, but the 2-inch shale in the Cathedral dolomite on Castle Mountain is no more a formation than is the 15-foot limestone in the upper part of the Comet shale 1200 miles to the south. To correlate the Cathedral dolomite in Canada with the Comet shale in Nevada is largely meaningless, if correlation means that the two formations were deposited at the same time and, therefore, are of the same age.

In an attempt to eliminate the confusion in correlation between faunal horizons or zones and formations the type of correlation table used by Bridge (1937, p. 234) was adopted (Deiss, 1938b, p. 1162). In this type of correlation chart (Fig. 5) the relative stratigraphic position of the successive faunas is indicated by bold-faced dots in each section in which they are known and the formations are adjusted to the faunas they contain. Thus fossils are the only basis of correlation, and the positions, and therefore the implied correlations of the unfossiliferous strata, are given solely in relation to the nearest faunal zone. On such a chart the stratigraphic sections are correlated, the temporal relationships of the formations are shown, but no one formation is correlated with any other.

LOCAL CORRELATIONS

The sections in the northern part of the area (Fig. 1) are similar lithologically except for the Ptarmigan and Stephen formations. The relationships of these formations are discussed in the earlier pages of this report. All the sections remeasured in 1938 and 1939 are compared and correlated

on Figure 4. The base of the *Kochaspis cecinna* zone and in its absence the base of the Cathedral dolomite is used as the datum plane for the different sections except in the Assiniboine area. This line is also the

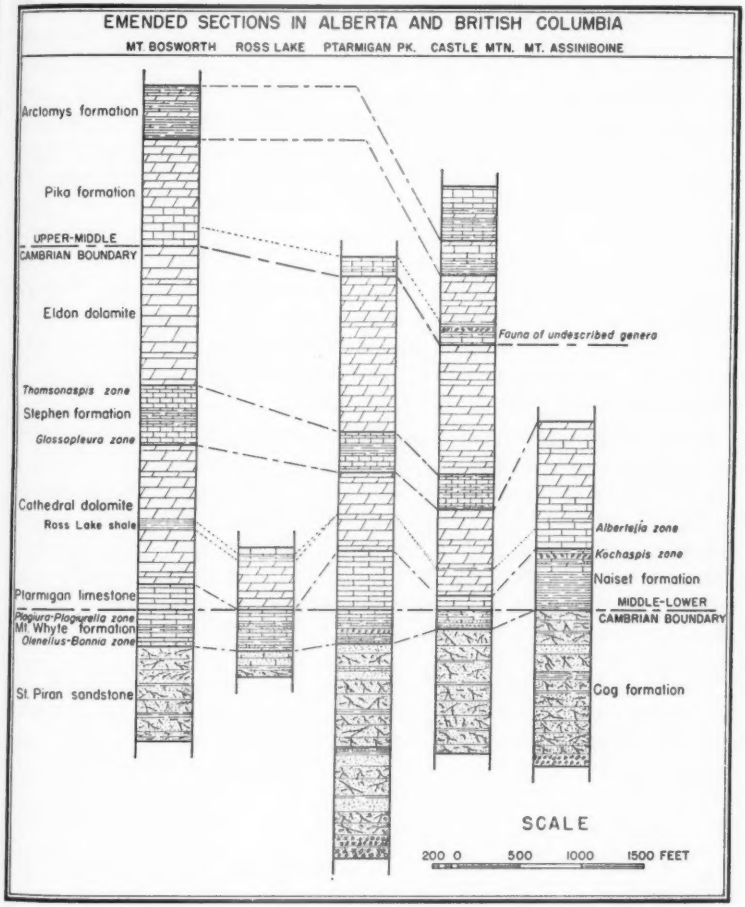

FIGURE 4.—*Relationships and correlation of formations in area*

Middle-Lower Cambrian boundary. In the Assiniboine section the Gog formation is tentatively considered the youngest Lower Cambrian formation. (*See* the discussion under "God sandstone.")

REGIONAL CORRELATIONS

General statement.—Precise correlation of the Cambrian sections must ultimately rest upon fossils and largely upon the trilobites. Consequently one of the most important purposes of the field work since 1932 in the Cordilleran trough has been to locate exactly the stratigraphic position of each Lower and Middle Cambrian faunule and to determine the vertical range of each genus. By 1937 eight faunal zones were tentatively established for the Middle Cambrian in the southern part of the Cordilleran trough (Deiss, 1938a, p. 274-275). The field work in 1938 and 1939 demonstrated that at least six of the eight faunas are also present in the Canadian Rockies. To these may be added two Lower Cambrian faunal zones which are present in both Nevada and Canada. (1) The *Olenellus-Bonnia* zone is limited to the basal beds of the Mount Whyte and the upper beds of the St. Piran formation. (2) The *Plagiura-Plagiurella* zone is 120 to 190 feet above the *Olenellus-Bonnia* zone on Ptarmigan Peak and Mount Bosworth respectively. Because much less field work has been done on the Waucobian than on the Albertan rocks and fossils in the Cordilleran region, the faunas are not so well studied or zoned as those from the Middle Cambrian.

Middle Cambrian faunal succession.—Sufficient field and laboratory evidence is now available to establish more precisely than before the Middle Cambrian faunal succession of trilobites in the Cordilleran region. As is to be expected in an area more than 1200 miles in length, all the faunas are not represented in one place, and geographic facies of each fauna are recognizable. The most complete and well-demarcated faunas occur in northwestern Montana (Deiss, 1939b, p. 63-130), but in that area the oldest Albertan fauna is absent.[8] The Middle Cambrian faunal sequence is:

(1) The oldest Middle Cambrian fauna in the Cordilleran trough occurs in the lower beds of the Comet shale (Deiss, 1938b, p. 1154) in Nevada and in the lower part of the Ptarmigan limestone in Alberta and British Columbia. In Nevada the fauna is characterized by *Kochaspis liliana* (Walcott), *K. augusta* (Walcott), *K. highlandensis* Resser, and *K. nevadensis* Resser, and several new genera. In Alberta and British Columbia the fauna consists of *Kochaspis cecinna* (Walcott), *K. gogensis* (Walcott), *K. chares* (Walcott), several other trilobites, and a few brachiopods.

(2) The second, or *Albertella,* fauna occurs 260 feet above the *Kochaspis* zone in Nevada and 580 feet above it on Mount Bosworth in British Columbia. The *Albertella* fauna as it is developed in the Ross Lake shale (Deiss, 1939a, p. 1004) seems to be restricted to the Canadian

[8] Resser's (1939, p. 11) *Ptarmigania* fauna of Utah and Idaho is a mixture of two widely separated Middle Cambrian faunas.

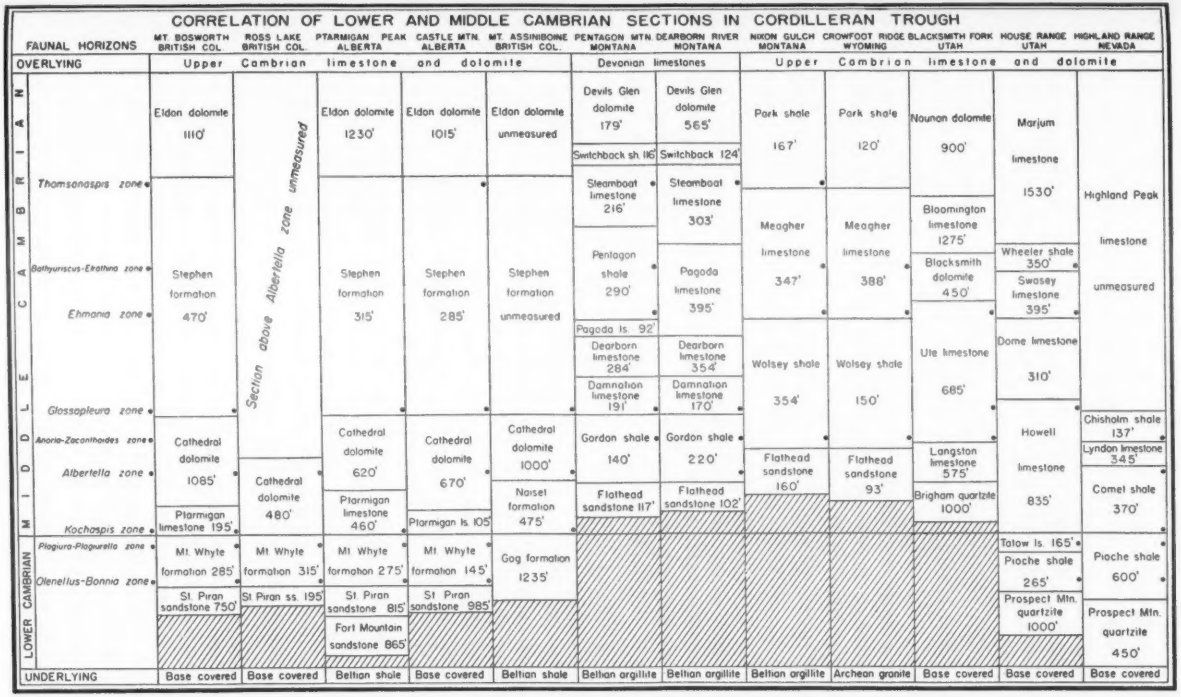

FIGURE 5.—*Regional correlation chart*

region, but *Albertella* and a few other genera are widespread and possess a short vertical range.

(3) The third Middle Cambrian assemblage, called the *Anoria-Zacanthoides* fauna (Howell and Mason, 1938, p. 296-297), is 275 feet stratigraphically above the *Albertella* zone in Nevada and was referred to as the *Zacanthoides cnopus* zone in Montana (Deiss, 1938a, p. 274; 1939b, p. 37-38) where the fauna contains many species that Walcott (1917b, p. 22-23) erroneously assigned to the *Albertella* fauna.

(4) The fourth fauna, *Glossopleura*, is probably the most widespread, narrowly restricted vertically, and one of the most valuable assemblages of trilobites in the Cordilleran trough. The *Glossopleura* fauna is present in the House Range and Blacksmith Fork, Utah, in northwestern Montana, and in every section remeasured in Canada. In northwestern Montana it occurs 30 feet above the *Anoria-Zacanthoides* zone and in Canada marks the basal 67 feet or less of the Stephen formation which is 470 to 700 feet stratigraphically above the *Albertella* zone. *Glossopleura* recurs in the base of the Dearborn formation in Montana 170 feet above its first occurrence in the Damnation limestone (Deiss, 1938a, p. 274). This recurrent fauna seems to be restricted to northwestern Montana and is no longer considered a separate faunal horizon elsewhere in the Cordillera.

(5) The fifth fauna, named from the trilobite *Ehmania* (Deiss, 1938a, p. 274), has never been found in the Canadian Rockies. The *Ehmania* fauna is well developed in northwestern Montana (Deiss, 1939b, p. 40-41) 570 feet above the *Glossopleura* zone and in the House Range 390 feet above the same zone.

(6) The sixth Middle Cambrian fauna always consists of numerous species and individuals wherever it is known but appears to be haphazardly distributed. This is the *Bathyuriscus-Elrathina* fauna which is widespread, but most prolific in the Pentagon shale of northwestern Montana (Deiss, 1939b, p. 44, 64-130). A facies of this fauna from the *Ogygopsis* shale was the first Cambrian biota described from the Canadian Rockies (Rominger, 1887, p. 12-19). The assemblage restricted to the Spence shale in Idaho is another facies of the *Bathyuriscus-Elrathina* fauna. The more common genera in this fauna are those usually given in general textbooks on historical geology as typical of the Middle Cambrian. Although *"Agnostus"* first appears a few feet below the *Bathyuriscus-Elrathina* zone in Montana, the genus is first greatly developed in the Cordilleran region in this fauna. In 1938 the writer first called attention to the oldest agnostids in the area by saying (Deiss, 1938b, p. 1163):

> "That '*Agnostus*' should first appear in the House Range in the basal beds of the Wheeler shale and in the Pentagon shale of Montana is significant, as in both areas this horizon is underlain by the *Ehmania* zone, and seems to indicate a correlation of the Pentagon and the Wheeler faunas."

The conclusion given seems to be essentially correct and is strongly supported by the fact that in every section studied in 1938 and 1939 in Canada "*Agnostus*" is absent except on Mount Stephen where it occurs in the *Ogygopsis* shale. Walcott's sections in Alberta and British Columbia support the conclusion that *Agnostus* did not reach the Cordilleran trough until *Bathyuriscus-Elrathina* time because the lowest Middle Cambrian beds in which he reported *Agnostus* are in the upper part of the Stephen on Mount Bosworth (Walcott, 1928, p. 312, 361) at least 325 feet above the *Glossopleura* zone, and in the Titkana formation on Robson Peak at least 1700 feet stratigraphically above the *Albertella* zone.

(7) The seventh and youngest group of trilobites known from the Middle Cambrian in the Cordilleran trough was originally named the *Kochaspis upis* fauna (Deiss, 1938a, p. 275). These trilobites were known only from northwestern Montana until 1938 when *Thomsonaspis*, one of the diagnostic genera, was found in the upper part of the Stephen formation on Castle Mountain (Deiss, 1939a, p. 967, 1005, 1021). Because *Thomsonaspis* occurs in Alberta and Montana and is useful in correlation between these areas, this biota is named the *Thomsonaspis* fauna. In Montana the fauna is in strata 225 to 400 feet stratigraphically above the *Bathyuriscus-Elrathina* zone and is commensurately younger than the *Ogygopsis* fauna of Mount Stephen. The other important genera of the *Thomsonaspis* assemblage are *Coelaspis, Glossocoryphus, Mcnairia,* and aberrant species of *Glyphaspis* and *Kochaspis* (Deiss, 1939b, p. 45).

The regional correlation chart (Fig. 5) represents the 12 sections now considered important in the Lower and Middle Cambrian stratigraphy of the Cordilleran region. The sections are arranged geographically from north to south. The faunas known from each section are indicated by a dot on the same plane as the name of the fauna.

PALEOGEOGRAPHY AND SEDIMENTATION

GENERAL STATEMENT

The principal discussion of the Cambrian geography and sedimentation in the southern Canadian Rocky Mountain region was by Walcott (1927, p. 148-172), who postulated five subsidiary troughs within the Cordilleran geosyncline during pre-Devonian time. The picture presented seems unnecessarily complex and is largely incorrect for the Lower and Middle Cambrian epochs because the conclusions reached are based upon too broad generalizations and insufficient or incorrect field data. When the area has been adequately mapped, the apparent haphazard distribution of the pre-Devonian sediments probably can be more rationally explained as the result of post-Paleozoic orogeny and subsequent erosion.

The following discussion of the paleogeography and sedimentation in

the central part of the Cordilleran trough is limited solely to the Cambrian period and largely to the Waucobian and Albertan epochs.

MONTANA ISLAND

One of the most important geographic elements in the Cordilleran trough during the Cambrian, Ordovician, and Silurian periods was the positive area which Walcott (1908b, p. 191) originally referred to as the Kintla uplift and later described as the Montana Island (Walcott, 1915, p. 197-198). The importance of this landmass (Fig. 6) has become increasingly apparent during the past 6 years as additional stratigraphic sections were measured and the distribution of the Cambrian formations and faunas became understood. Few, if any, stratigraphers have realized the influence of this landmass on the character of the contiguous Lower and Middle Cambrian sediments and upon the distribution of the Cambrian, Ordovician, and Silurian rocks and fossils within the 50,000 square miles or more which lie between Great Salt Lake, Utah, and southern British Columbia and Alberta. The field data now available indicate that the Montana Island during the Early Paleozoic in the central part of the Cordilleran trough may have been as important as the Cincinnati-Nashville anticline during the Late Ordovician to Pennsylvanian periods in eastern United States. The role played by the Montana Island was discovered in connection with the Pentagon shale problem in 1933, and the effects of the landmass on Cambrian sedimentation in Montana were first discussed in 1939 (Deiss, 1939b, p. 58-60, Fig. 7).

The field work in Alberta and British Columbia in 1938 and 1939 and the discoveries of Cambrian fossils in isolated areas north and west of Montana by other workers disclosed sufficient new evidence to warrant description of the broader outlines and effects of the Montana Island during Lower and Middle Cambrian time in southeastern British Columbia and southwestern Alberta.

The use of the term Montana Island creates a somewhat confusing and erroneous picture because the landmass was rarely an island. Instead, it was more nearly a peninsula at times and a complete east-west land barrier at others. Probably the mass was entirely surrounded by water only during part of the Middle Cambrian.

LOWER CAMBRIAN SEAWAYS

The most northern locality in southwestern United States from which Lower Cambrian fossils have been reported is on Promontory Point (Resser, 1939, p. 6), the peninsula on the north side of Great Salt Lake, Utah. The most southern Waucobian fossil locality in the Canadian Rockies is 6 miles northeast of Cranbrook, British Columbia (Schofield, 1922, p. 9), approximately 40 miles due north of the International

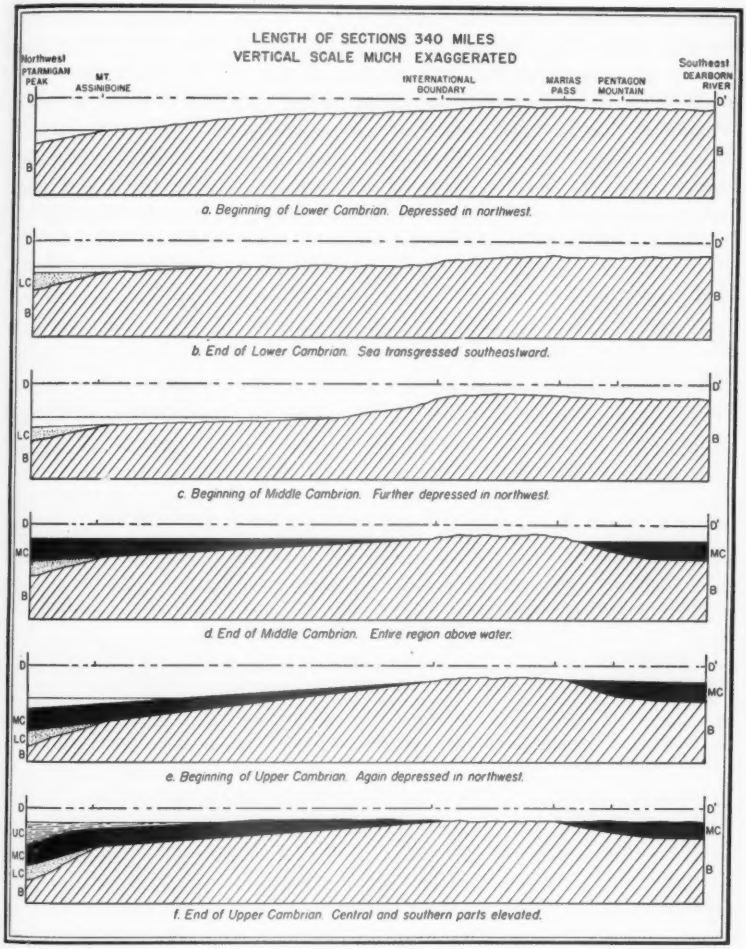

FIGURE 6.—*Diagrammatic sections to illustrate Cambrian sedimentation in vicinity of Montana Island*

D D′ = horizontal datum plane, U C = Upper Cambrian sediments, M C = Middle Cambrian sediments, L C = Lower Cambrian sediments, B = Beltian; blank areas indicate positions of seas at beginning of epochs.

Boundary. Branson (1931, p. 70) reported *Kutorgina cingulata* (Billings) from a locality on the Colville River in northeastern Washington, and Resser (1934, p. 7) identified from the same general region ". . . frag-

ments that appear to represent *Wanneria*, or at least an olenellid trilobite, accompanied by several cups of *Archaeocyathus*." The writer examined the collection made by W. A. G. Bennett from the Colville locality [9] and is able to confirm Resser's conclusion. Aside from the Colville locality Waucobian fossils have never been found between Cranbrook, British Columbia, and Promontory Point, Utah, and there is no evidence that the Lower Cambrian sea ever invaded Montana and Idaho, or that marine waters were continuous from Canada to Utah during the Waucobian epoch.

In discussing the Cambrian rocks of the Wasatch Mountains Resser (1939, p. 6) said:

> "In my opinion the Lower Cambrian rocks found in the vicinity of Great Salt Lake possibly continue beneath younger strata and the lava fields, to connect with outcrops in northeastern Washington and beyond that the beds of similar age in the Columbia Valley and Dogtooth Mountains of British Columbia."

This statement was largely motivated by the fact that the faunal sequence in the Mount Whyte, Ptarmigan, and Cathedral formations near Kicking Horse Pass (Fig. 1) is astonishingly similar to that in the Pioche and overlying Comet shale in the Highland Range of Nevada. That Walcott also believed the Cordilleran trough was continuous between Canada and the southern United States during Waucobian time is shown in his discussion of the Bow trough in Alberta and British Columbia (Walcott, 1927, p. 162):

> "The seaway of the Trough was probably from 50 to 60 miles (80.5 to 96.5 km.) in width, at the time of its greatest development. Its extension to the south-southwest appears to have been limited by the pre-Cambrian Kintla Island uplift, but it was undoubtedly connected on the south with an open seaway, as yet unrecognized, of the Cordilleran Geosyncline, for the Lower and Middle Cambrian faunas of Utah and Nevada are closely related to those of the Bow Trough Lower and Middle Cambrian formations."

Similar Lower Cambrian trilobites do occur in the Canadian Rockies and in Utah, and part of the *Olenellus* fauna reached as far south as northeastern Washington, but all the other field evidence suggests that two separate invasions occurred in the Cordilleran geosyncline in Waucobian time: one from the south through southern California, and the other from the north through Alaska. The southern sea extended northeastward to northern Utah, and the northern sea extended southward to the 50th parallel near Mount Assiniboine and to the southwest beyond the 49th parallel near Colville, Washington.

Waucobian marine waters entered the northern part of the Cordilleran trough and transgressed southward over the gently undulating eroded surface of Beltian sediments (Fig. 6, a). Sands and muds accumulated

[9] Through the courtesy of Dr. Harold E. Culver, Supervisor of the Division of Geology, Washington Department of Conservation and Development, the collection was sent to the writer. The trilobites are extremely fragmentary, but several broken cheeks possess a similar surface ornamentation to that of *Wanneria walcottanus* (Wanner).

in the vicinity of Ptarmigan Peak (Fig. 1) on the east side of the trough to a thickness of at least 1620 feet (Deiss, 1939a, p. 980-985). These clastic sediments thinned southward toward the Montana Island where gravel, sand, and mud were deposited long after the seas had cleared in the north. During the last phases of clastic deposition in the northern area (oldest Mount Whyte time) individuals of the *Olenellus-Bonnia* fauna first arrived in the area and migrated southeast possibly as far as Mount Assiniboine. These trilobites followed the northwest shore of the Montana Island barrier to the vicinity of Colville, Washington, approximately 30 miles south of the International Boundary.

The transgression was halted on the north shore of the Montana Island where sands and gravels (Gog formation) were being deposited during late Waucobian time while limestones and shales (Mount Whyte formation) were being deposited in the north. The evidence for this statement is the much older and thicker series of basal Cambrian clastic sediments at Ptarmigan Peak than are present near Mount Assiniboine and southward. Likewise the Mount Whyte formation thins from 275 feet at Ptarmigan Peak to 145 feet at Castle Mountain (Deiss, 1939a, p. 970-971, 980-982) and is absent or represented by the upper part of the Gog sandstone at Mount Towers 30 miles to the southeast. These facts combined with the absence of any known Waucobian deposits throughout Montana and Idaho strongly support the hypothesis that the central part of the Cordilleran trough (Montana Island) was a low-lying landmass over which the northern Waucobian sea transgressed as far south as Colville, Washington.

If, as postulated, the northern and southern parts of the Cordilleran geosyncline were separated throughout the Waucobian epoch, the presence of similar faunas in the two regions on opposite sides of the land barrier would be accounted for by the migration of similar genera and species from a northern Pacific radiation center. From this point some individuals migrated along the west shore of Cascadia and thence into Nevada and Utah while others from the same parent fauna entered the northern part of the trough around the north end of Cascadia and then migrated southward with the transgressing northern sea.

EXPANSION OF MIDDLE CAMBRIAN SEAWAYS

By the opening of the Albertan epoch (Fig. 6, c) sands and gravels were still being deposited on the north side of the Montana Island. Erosion of the Beltian quartzites and argillites (Kintla and younger formations) of which the Island was formed furnished most of the larger pebbles and boulders in the basal conglomerate south of Mount Assiniboine. The northern waters were clear and in them was precipitated the older part of the Ptarmigan limestone and Cathedral dolomite, while the muds of

the Naiset formation were being deposited near Mount Assiniboine. Shortly thereafter (*Albertella* time) the southern sea transgressed over the south and west sides of the landmass in Montana and Idaho where the Flathead (Deiss, 1939b, p. 35-37) and Gold Creek (Sampson, 1938, p. 9-10) sands were deposited. About the same time the northern sea advanced nearly as far south as the International Boundary west of Waterton Lakes where gray-green and red shales resting on sandstone and overlain by limestones contain Middle Cambrian fossils (Hume, 1933, p. 6b). The lithology and sequence of these rocks closely parallel the Flathead, Gordon, and Damnation formations on the south side of the Montana Island in northwestern Montana (Deiss, 1939b, p. 52).

Slightly later in the Albertan the northern and southern waters became confluent on the west side of the Montana Island in the vicinity of Metaline Falls, Washington. These seas may have merged earlier in the epoch, but the oldest fossils from Metaline Falls are of approximately the same age as the *Bathyuriscus-Elrathina* fauna in northwestern Montana. The waters to the north and south of the island remained clear during most of late Albertan time and in them were deposited the thick limestones, shales, and dolomites of the Stephen and Eldon formations to the north and the Pentagon Steamboat Switchback and Devils Glen to the south (Fig. 6, d).

By the beginning of Upper Cambrian (Fig. 6, e) the central and western parts of the Montana Island were re-elevated. The Cordilleran sea spread southeast through Alberta and north-central Montana, and thence south and west around the southern shore of the Montana Island. Consequently, Croixan sediments were not deposited in northern Idaho, northwestern Montana, and extreme southeastern British Columbia (Fig. 6, f).

CONCLUSIONS

The field work in Alberta and British Columbia and the study of the literature and trilobites seem to justify the following conclusions:

On Mount Bosworth the *Albertella* (Ross Lake) shale is well exposed and extremely fossiliferous on the east end of the mountain in the Cathedral dolomite 385 feet stratigraphically above the top of the Ptarmigan limestone. The Eldon dolomite is normal in thickness, being 1110 instead of 2728 feet as originally stated. The rocks which form the top of Mount Bosworth belong to the Pika instead of either the Arctomys or Bosworth formations which, however, are present higher in the section. Only 750 feet of St. Piran is exposed instead of 2705 feet, and the Lake Louise and Fort Mountain formations are not accessible. The total thickness of Cambrian strata in the Mount Bosworth section is more nearly 6300 than 12,353 feet as originally reported.

The *Ogygopsis* shale on Mount Stephen seems to rest on the Cathedral

instead of beneath the Eldon and thus forms the base instead of the upper part of the Stephen formation in its type locality. The solution of the Stephen problem remains partly unsolved. Conclusive evidence for its solution will be difficult to obtain as the rocks on Mount Stephen are much deformed and faulted and the *Ogygopsis* shale may not be present in the unexplored adjacent sections to the east.

The Mount Whyte formation is not represented near Mount Assiniboine, and the rocks which Walcott assigned to the formation contain *Kochaspis*, *"Amecephalus"*, and other Middle Cambrian fossils. These strata are described as a new formation—the Naiset.

Albertella is present in the Cathedral limestone on Naiset Point (Fig. 1) and makes certain the correlation of the beds there with the Cathedral formation (Fig. 4) in the sections to the north.

The Fort Mountain, St. Piran, and Mount Whyte formations thin southward between Ptarmigan Peak and the Assiniboine region to the extent that the series is doubtfully represented by only 1235 feet of quartzites, sandstones, and shales, all of which are probably late Waucobian in age. The basal Cambrian clastic sediments in the Assiniboine region, therefore, probably are correlative with the Mount Whyte and St. Piran formations and are assigned to the newly described Gog formation.

The Lower and Middle Cambrian formations can be correlated and traced within the northern part of the area but as formations they are not precisely correlative with any formations in Montana, Utah, or Nevada. The stratigraphic sections in Alberta and British Columbia are correlated with seven sections in western United States (Fig. 5) on the basis of widely distributed faunas each of which possesses a relatively short vertical range.

The Lower and Middle Cambrian faunal succession is described for nine assemblages of trilobites whose vertical range and geographic distribution are relatively well known. Two of the faunas are restricted to the Waucobian (1) *Olenellus-Bonnia*, (2) *Plagiura-Plagiurella*, and seven are in the Albertan (3) *Kochaspis*, (4) *Albertella*, (5) *Anoria-Zacanthoides*, (6) *Glossopleura*, (7) *Ehmania*, (8) *Bathyuriscus-Elrathina*, and (9) *Thomsonaspis*.

"Agnostus" probably first entered the Cordilleran trough in late Middle Cambrian time. At least the genus (or group) has never been found in rocks older than the Wheeler and upper Ute in Utah, the Pentagon and Meagher in Montana, and the upper Stephen in British Columbia.

Cambrian geography and sedimentation in southern Alberta and British Columbia were much influenced by a landmass whose apex lay in the area occupied today by Glacier National Park (United States). This landmass—the Montana Island—probably extended from Colville, Washing-

ton, to northern Utah during the Lower Cambrian, was surrounded by water in the Middle Cambrian, and was re-elevated in its western and central parts before and during Upper Cambrian. As a result of the Montana Island probably 550 miles of the central Cordilleran Geosyncline was land during Waucobian time when the seas entered separately from the Arctic and through southern California. The Cordilleran trough was continuous from British Columbia to Montana through northern Idaho and northeastern Washington on the west side of the Montana Island during the Middle Cambrian. The western part of the trough was again above sea level in northwestern Montana, Idaho, Washington, and southeastern British Columbia in the Upper Cambrian when the Cordilleran sea spread southeast through Alberta and north-central Montana and thence south and west around the southern shore of the Montana Island.

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